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MECHANISMS OF PEGMATITE EMPLACEMENT

By ROBERT A. CHADWICK

ABSTRACT

Various mechanisms by which pegmatites may have come into place are considered, and the following classification of modes of emplacement, based largely on structural concepts, is suggested:

Displacement: country rocks moved aside from site of emplacement
  Forceful emplacement: country rocks moved aside by pegmatite
  Permissive emplacement: pre-existing opening in country rock filled by pegmatite
Nondisplacement: no movement aside of country rocks occurred
  Mobility: material at pegmatitic site was bodily mobile during emplacement
    Introduction: pegmatitic material introduced into site of emplacement
    Local derivation: pegmatitic material derived locally
      Host-rock breakdown: pegmatite derived by chemical breakdown of pre-existing rock
      Residual segregation: pegmatite derived by crystallization, in place, of residual magmatic fluids
  Nonmobility: material at pegmatitic site not bodily mobile during emplacement
    Introduction
    Local derivation

Criteria that may help to distinguish these various modes of emplacement are listed and discussed. Particular emphasis is given to field features. In the literature disagreement exists concerning the worth of some of the criteria, whereas there is substantial agreement on other lines of evidence.

Results of a field study of certain granitic pegmatites of New England are also presented. The purpose of this study was to determine the modes of emplacement of these bodies by application of the various criteria developed. The mode or modes of emplacement of some of these pegmatites could be partly determined, as the bodies exhibit critical features. The modes of emplacement of other bodies were difficult to determine because critical features are lacking.

The various pegmatites developed by several mechanisms. Furthermore, individual pegmatites commonly exhibit evidence for more than one process. Country rock was displaced forcefully to make room for some pegmatites or portions of them and was displaced either forcefully or permissively at other bodies. At some pegmatites the country rock was not bodily displaced, and the material at the sites of most of these bodies was mobile. Pegmatite matter was introduced into emplacement sites where mobile or nonmobile processes were active.

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Introduction

Purpose and Scope of Study

The question of mode of emplacement is a major aspect of the general problem of pegmatite origin. If the mechanisms by which pegmatites are developed in their present sites could be determined, much light would be shed on related problems such as the source of pegmatitic constituents, their transport to the sites of emplacement, and the post-crystallization history of pegmatite bodies.

Various mechanisms of emplacement have been proposed in the literature for specific pegmatites or groups of pegmatites, but no comprehensive discussion of the subject has been presented. The writer believes that structural concepts may serve as a basis for a systematic analysis of the emplacement problem. He attempted first to classify possible modes of emplacement; second, to assemble ideas on emplacement from the literature; and third, to list and evaluate, in so far as possible, the criteria of emplacement that have been suggested. Particular attention has been given to field criteria. Finally, the writer studied a

number of well-exposed granitic pegmatites in New England in the field and by applying the best of the criteria attempted to determine their mode or modes of emplacement.

Previous Work

Most early workers believed that pegmatites are emplaced by an invasion of fluid, either essentially magmatic or aqueous. Kemp (1924) summarized many of the early views. Andersen (1931) discussed structural features in and around pegmatite bodies and attempted to relate them to emplacement. He concluded that many pegmatites crystallize by forcibly intruding the country rock or by permissively draining into structural openings. Chapman (1941), largely on the basis of structural evidence, concluded that some New Hampshire pegmatites result from segregations in magmas and some from the inflow of pegmatitic fluids into fractures in a crystal mush or a solid rock. Uspensky (1943) believed that partial recrystallization of aplite is a factor in the development of some bodies. Quirke and Kremers (1943) suggested that pegmatites can form either by filling of open space or by metasomatic replacement, depending on the speed of flow of pegmatitic fluids in an open channel.

Recently, detailed studies of zoned pegmatites convinced many workers that these bodies result primarily from fractional crystallization of pegmatitic fluid inward into a chamber to which no pegmatitic solutions were added while the fractional crystallization took place. Among these workers are: Cameron, Jahns, McNair, and Page (1949); Flawn (1951); Jahns (1952); Jahns, Griffitts, and Heinrich (1952); Page *et al.* (1953). Some of these authors also noted structural features in the wall rocks around certain pegmatites and used these features as evidence for various modes of emplacement—processes that involve displacement of the host rocks and those that do not (Jahns, 1946; Cameron, 1951; Jahns, 1952; Jahns, Griffitts, and Heinrich, 1952; Page *et al.*, 1953; Cameron *et al.*, 1954).

Several authors (King, 1948; Higazy, 1949; Ramberg, 1949; 1952; 1956; Barth, 1952) recently stressed metasomatic replacement, metamorphic differentiation, recrystallization, concretion (crystals growing outward from a center in a solid medium, pushing aside the wall rocks), and secretion (diffusion of particles to openings, where they crystallize into open space) as modes of pegmatite emplacement.

A review of theories of pegmatite origin, including emplacement, was recently provided by Jahns (1955).

Acknowledgments

The writer's study was made under the direction of Dr. E. N. Cameron, who first suggested the problem, furnished maps and information, and critically read the manuscript. This help is gratefully acknowledged. Sincere thanks are also due the other staff members of the Geology Department of the University of Wisconsin, who made the facilities of the Department available to the writer. The pegmatite mine owners and operators co-operated in many ways. R. L. Coons ably assisted with the field work during the summer of 1954. Mrs. Diane K. Chadwick was of invaluable assistance at various stages of preparation of the manuscript and in drafting the illustrations.

Definition of Pegmatite Used

The term pegmatite is used in this paper in the sense of Andersen (1931, p. 3) who said,

> "Pegmatites are mineral associations crystallized in situ, decidedly more coarse-grained than similar mineral associations in the form of ordinary rocks and differing from these in having a more irregular fabric of the mineral aggregates."

The present discussion is limited to pegmatites that are essentially granitic, that is, composed chiefly of quartz and alkali feldspar. This definition is intended to avoid genetic implications in so far as possible.

Pegmatite, as used in the text, may refer to the rock type itself or to a body of pegmatite rock.

Possible Mechanisms of Emplacement

The writer attempted to delimit and classify possible mechanisms of emplacement of pegmatites. Under the system adopted, the classification of a pegmatite proceeds through one or more of a series of stages; two processes are contrasted at each stage. The method involved is brought out by a series of questions about the emplacement of a given pegmatite. Most of these questions are based on structural and textural features of the pegmatites and the surrounding wall rocks:

(1) Was the country rock displaced to make room for the pegmatite?

(2) Assuming displacement, was the country rock moved aside by force operating from the

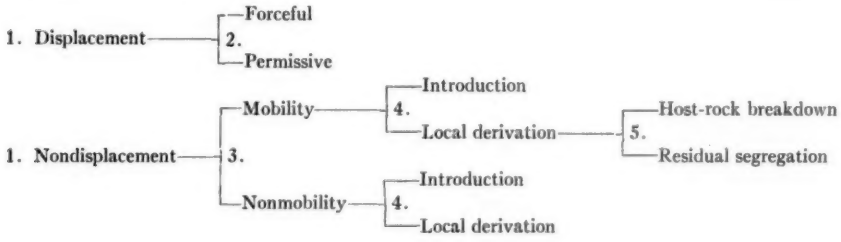
pegmatite, or by a force operating from the outside which produced an opening for the pegmatite to enter permissively?

(3) Assuming lack of displacement, did the material occupying the site of emplacement become bodily mobile during the emplacing process, causing mass flowage or movement, or was it metamorphically or metasomatically transformed in place into pegmatite with possible preservation of former texture and structure?

(4) Were the pegmatitic constituents introduced into the site of emplacement, or were they locally derived?

(5) Assuming local derivation, did the pegmatite develop by the breakdown of solid host rock or did it crystallize from a magma as a residual fluid segregation?

The modes of emplacement proposed in these questions are outlined as follows:



Definitions of Mechanisms of Emplacement

Displacement

Definition and subtypes.—The rock formerly occupying the site of the present pegmatite was moved bodily outward either before or during emplacement to make room for the pegmatite. This displacement was either forceful or permissive.

FORCEFUL EMPLACEMENT: Growing or crystallizing pegmatitic material forced aside the walls of an opening which was perhaps extremely small. This force was either pressure acting from an intrusive or locally derived fluid or force exerted by mineral aggregates growing crystoblastically in an essentially solid medium and bodily pushing aside the host rock. Regional deformation contemporaneous with emplacement is considered forceful to the extent that the pegmatitic material moved under large-scale forces, exerted pressure upon the surrounding rock, and deformed it. The deformation may have occurred during only the final stages of emplacement of a pegmatite which had developed by other processes before that time; however, the classification is necessarily based on the stage or stages from which geologic evidence is preserved, perhaps only the final one.

PERMISSIVE EMPLACEMENT: The pegmatite crystallized passively in an opening formed before or during the ingress of pegmatitic material consisting perhaps of fluid or perhaps of constituents diffusing without the movement of a pegmatitic fluid—similar to the "secretion" mechanism of Ramberg (1952, p. 252).

Nondisplacement

Definition and major subtypes.—Processes that do not involve bodily displacement of the host material from the site of the pegmatite may be termed types of nondisplacement.

MOBILITY: Material that occupied the site of the present pegmatite, either the original country rock or some product formed during emplacement, became bodily mobile at some time during the emplacing process, so that mass flowage or movement took place. The presence of a liquid or gas of pegmatitic composition is not necessarily implied. Rather than emphasize the condition of the medium from which the pegmatite developed, the definition is intended to include all processes that could produce certain observable effects—those consequent upon mass movement of the material present. Examples of mobile processes might include stoping of host rock, rheomorphism consequent on metasomatic transformation, mass melting or solution, and plastic flowage or other localized movement during emplacement that would produce rotation or translation, en masse, of included fragments of the country rock. Solid host rock need not be present; residual crystallization of pegmatite from a larger magmatic body entails mobility of the material occupying the pegmatitic site before crystallization of the body.

NONMOBILITY: The material at the site of emplacement was never mobile en masse but rather was dissolved and recrystallized or replaced by an intimate, increment-by-increment process. Outlines of the pre-existing texture and structure may or may not be preserved in the resulting pegmatite. Nonmobility might be effected by recrystallization, crystalloblastic growth in which the host rock is not bodily pushed aside from the site of emplacement, metamorphic differentiation with the same qualification, or metasomatism.

Subclassifications of mobility and nonmobility.—Both mobility and nonmobility can be subdivided according to the source of the pegmatite-forming material.

INTRODUCTION: At least some of the pegmatitic constituents came into the site of emplacement from elsewhere. Possible mechanisms of introduction are: stoping into a chamber below the site of the present pegmatite, assimilation of wall rock by introduced pegmatitic material, and metasomatic transformation of country rock.

LOCAL DERIVATION: The material composing the pegmatite was derived essentially at the site of emplacement; any solid host rock present was dissolved, differentiated, or transformed without additions from beyond the immediately surrounding rocks. If a mobile process, local derivation can be subclassified according to whether the body was derived from pre-existing solid rock (host-rock breakdown), or whether no solid host rock was present (residual segregation). A residually segregated pegmatite is one that crystallized from locally derived fluids remaining during the solidification of a larger magmatic body. Residual fluid that moved a considerable distance through a shifting, crystallizing mush before emplacement is considered permissively emplaced.

Additional Comments

Most of the processes defined above can take place either in an open system, where pegmatitic material is added to the site of emplacement during the primary crystallization of the body, or in a restricted system, where it is not. The degree of restriction of the system is not a principal consideration in the above classification. Mechanisms operative after completion of the primary crystallization of the body—hydrothermal replacement, for example—are outside the scope of the present study.

Natural modes of emplacement may, of course, be complex combinations of the processes somewhat arbitrarily defined above. Some pegmatites may have developed by several mechanisms, and the features produced by one process might be changed by a succeeding one. At such bodies perhaps only the final stage in the emplacing process can be determined.

Discussion of Criteria

General Remarks

Various criteria that favor one mode of emplacement or another are grouped below into tables that contrast these processes. Each criterion is discussed individually. Some of these criteria have been considered at length in the literature, not necessarily with specific reference to pegmatites. Those that are believed to apply to pegmatites are included. The writer proposes some other criteria which to his knowledge have not been discussed in the literature, at least with reference to the mechanisms of emplacement they are here considered to indicate.

A criterion, as defined in this paper, need not signify necessary and sufficient proof for a given process but rather constitutes evidence favoring that mechanism over a contrasting process. Some lines of evidence are more reliable than others. The criteria near the top of each table are considered by the writer to be in general more reliable than those near the bottom. In the discussions following the tables, reasons are given for considering a feature relatively reliable or unreliable, but each reader should exercise his judgment as to the relative worth of the features listed. The original references should be consulted for a more detailed treatment.

Any conclusions about the mode of emplacement of a pegmatite should be considered valid for only the part or parts of the body to which the criteria apply; the body may have developed by more than one mechanism.

The numbering of the criteria tables corresponds to that of the questions posed at the beginning of the discussion. Each table presents two contrasting mechanisms of emplacement. Each criterion should be used only to distinguish between the modes of emplacement on the specific table on which it occurs, and the tables should be considered in numerical order. Criteria from Table 1 establish whether or not a pegmatite has displaced the host rock; those from Table 2 or 3 may then be used to determine the type of displacement or nondisplacement; criteria from Tables 4 and 5 classify the emplacing process into more detailed categories.

TABLE 1.—CRITERIA PERTAINING TO DISPLACEMENT AND NONDISPLACEMENT

Displacement	Nondisplacement
101. No units of wall-rock sequence are missing at site of pegmatite.	107. Unit in wall-rock sequence is missing; its site is occupied by pegmatite.
102. Walls of irregular pegmatite match.	108. Evidence indicates that emplacement was nonmobile.
103. Foliation of wall rock is deflected to parallel pegmatite contact. Evidence indicates that deformation is not post-emplacement.	109. Irregular, rounded pegmatite contact cuts across undisturbed wall-rock foliation.
104. Fracturing, bending, or folding is present in wall rock, concentrated around pegmatite margins. Evidence indicates that deformation is not post-emplacement.	110. Tabular units transected by pegmatite are narrower or absent on one side of body; feature is not due to faulting horizontally beyond exposure.
105. Obliquely transected wall-rock units are consistently offset directly across pegmatite.	111. Obliquely transected wall-rock units are consistently present on opposite sides of pegmatite without offset.
106. Regularly shaped units of wall rock increase proportionately in thickness where pegmatites occur within the unit.	112. Composition within a pegmatite varies with that of adjacent wall rock.

TABLE 2.—CRITERIA PERTAINING TO FORCEFUL AND PERMISSIVE EMPLACEMENT

Forceful	Permissive
201. Pegmatite shape is too irregular to reflect shape of unmodified pre-emplacement opening. Evidence for displacement, such as conformance of wall-rock foliation to irregularities of the contact, is also present.	204. Pegmatite follows regional tension fracture pattern.
202. Opposite pattern of drag folds or same direction of bending of foliation occurs in wall rocks on opposite sides of pegmatite. Evidence indicates that disturbance is not post-emplacement.	205. Pegmatite follows pattern of regional brecciated shear zones.
203. Wall rocks thrust outward from pegmatite. Evidence indicates that disturbance is not post-emplacement.	

TABLE 3.—CRITERIA PERTAINING TO MOBILITY AND NONMOBILITY

Mobility	Nonmobility
301. Primary flow or deformation structure is present in pegmatite.	308. Pegmatite contains relict textures or structures of the pre-existing rock.
302. Foliation in pegmatite is consistently parallel to contacts but nonparallel to wall-rock foliation.	309. Foliation in pegmatite is parallel to that of wall rocks across discordant contact. Foliation does not extend across entire pegmatite.
303. Foliation and lineation of inclusions in unfoliated pegmatite is nonparallel to that of wall rock or other inclusions. Surrounding wall rock is not disturbed enough to mask effects of any movement of inclusions.	310. Foliation and lineation of inclusions in unfoliated pegmatite is parallel to that of wall rocks.
304. Relation of fracture patterns to foliation or lineation in pegmatite suggests flow.	
305. Pegmatite is internally zoned.	
306. Pegmatite minerals are oriented perpendicular to contacts and widen inward from them.	
307. Inclusions are dissimilar in rock type to adjacent walls.	

The first digit of the number assigned to a criterion indicates the table in which it is found. The numbers of the criteria discussed in each paragraph or subsection below are noted in parentheses at the beginning of each of the subsections.

TABLE 4.—CRITERIA PERTAINING TO INTRODUCTION AND LOCAL DERIVATION

Introduction	Local derivation
401. Surrounding wall rock (*e.g.*, some quartzites or marbles) has chemical composition which eliminates it as sole source for pegmatitic constituents.	404. Adjacent wall rock is impoverished in pegmatitic constituents as compared to that farther from pegmatite.
402. Adjacent wall rock is enriched in pegmatitic constituents as compared to rock farther from pegmatite.	
403. High percentage of area is occupied by pegmatites, and wall rocks differ too much chemically from pegmatites to be sole source of pegmatitic material present.	

TABLE 5.—CRITERIA PERTAINING TO HOST-ROCK BREAKDOWN AND RESIDUAL SEGREGATION

Host-rock breakdown	Residual segregation
501. Host rock is sedimentary or metasedimentary.	503. Pegmatite is replaced or transected by material contemporaneous with crystallization of surrounding host rocks.
502. Evidence establishes that pegmatite is appreciably younger than host rock (*e.g.*, occurrence of an event between consolidation of the country rock and pegmatite emplacement).	504. Primary flow structure in host wall rock is modified by pegmatite.
	505. Protoclastic textures and structures in pegmatite can be correlated with late-stage structures in host rock.

Criteria Pertaining to Displacement and Nondisplacement

Evidence for nonmobility (*108*).—Features that are criteria for nonmobile processes such as recrystallization or replacement establish that the pre-existing rock was not displaced (108).

Presence or absence of country-rock units (*101, 106, 107*).—In many areas, authors have established by means of marker units such as distinct beds of the country rock that the entire sequence of rock wraps around an emplaced body: none is missing (101). Such evidence is generally considered to indicate displacement of the walls of the body (*e.g.*, Jahns, 1952, p. 55; Walton, 1955, p. 8).

Though distinctive marker zones may not be present, a large unit of country rock which is consistent in thickness over a long distance may increase in thickness in the vicinity of an emplaced body or bodies (106). Kruger (1946a, p. 201) considers that this feature favors displacement. Provided the increase in width is proportional to the amount of emplaced pegmatite, the writer considers that the feature suggests displacement; the argument is stronger if, as in Kruger's example, the feature is found on a small scale as well as large. Possible volume changes in the third dimension must always be considered.

According to many authors, the sudden disappearance at the site of emplacement of regular, persistent country-rock units indicates lack of displacement (107). Lack of displacement need not imply metasomatism, as the country rock may have been bodily melted or dissolved in place, assimilated, or stoped through the action of introduced material.

Offset of obliquely transected structures (*105, 111*).—A tabular unit in the country rock such as a bed or dike may be obliquely transected by a tabular pegmatite. If displacement of the country rock made room for the pegmatite, the older marker zone should be offset directly across the body at right angles to its trend, as

in Figure 1, *A* (105). Billings (1925, p. 145), Goodspeed (1940, Fig. 14), and King (1948, Fig. 1) emphasized the value of this feature if properly applied. If the walls were not displaced, the older zone should not be offset, as in Figure 1, *B* (111).

(1954, p. 105) believe that in an area of many intersecting fissures that outline blocks of wall rock, the pattern in which the fractures open would not be one of simple right-angle dilation. The movement on each fissure would be controlled by the shape and position of near-by

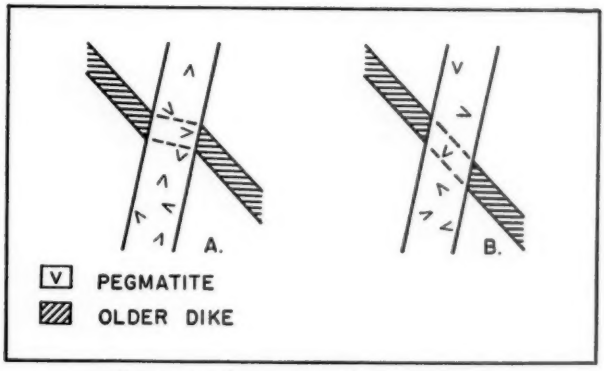


FIGURE 1.—PRINCIPLE OF DILATANT OFFSET
Pegmatite transects dike obliquely. (A) Walls separated to make room for pegmatite, older dike is offset. (B) Walls not separated, no offset occurs.

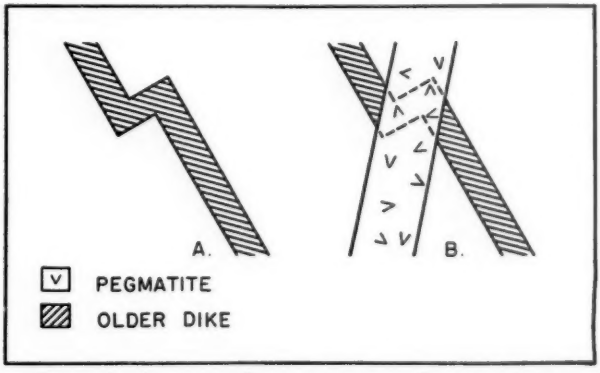


FIGURE 2.—ALTERNATE INTERPRETATION OF DILATANT OFFSET
(A) Older dike has sharp jog. (B) Pegmatite then emplaced without displacing older rock, but appearance is that of dilatant offset. (*Cf.* Fig. 1 *A*.)

If the above criterion is to be valid, the following must be assumed: (1) No movement in the plane of the fissure occurred either before, during, or after emplacement of the pegmatite along it; rather, all the movement of the country rock was at right angles to this plane. (2) The marker zone contained no sharp jogs or bends which were later masked by the pegmatitic material (Fig. 2).

Fissures need not dilate at right angles under regional tensional conditions. Wells and Bishop

blocks and might not be easy to predict. Separation of the walls could therefore take place without resulting in offsetting of the marker zone directly across the opening.

A strong argument against movement parallel to the plane of the fissure is provided if the emplaced body transects two or more tabular zones trending in different directions on the surface of exposure and dipping in directions other than perpendicular to this surface, and if there is consistent right-angle offset or lack of

any offset of these older zones (Fig. 3, *A*; *see* Billings, 1925, Fig. 5; Wells and Bishop, 1954, p. 106; Ramberg, 1956, p. 200).

Change in width of older zone across pegmatite (110).—A fairly regular marker zone in the country rock which is transected by a dilatant opening should measure the same width on both walls of any body that filled the opening. If the zone is narrower or even absent on one wall, (Fig. 3, *B*) the emplaced body did not move aside the older unit but perhaps metasomatically replaced or stoped it (*see* Billings, 1925, p. 146; King, 1948, p. 460–466.)

However, Wells and Bishop (1954, p. 106) demonstrate that if the transected older units are approximately equidimensional inclusions in the country rock, a slight component of movement along the fissure and perpendicular to the surface of exposure might "fault out" or cause to be narrower the portion of an inclusion on one side of the fracture (Fig. 4). Tabular zones of consistent thickness in the wall rock would not disappear or become narrower if faulted in this manner; thus only where the marker zone is tabular can its disappearance or change in width on one side of a pegmatite be considered reliable evidence for nondisplacement (110). In addition, the possibility must be

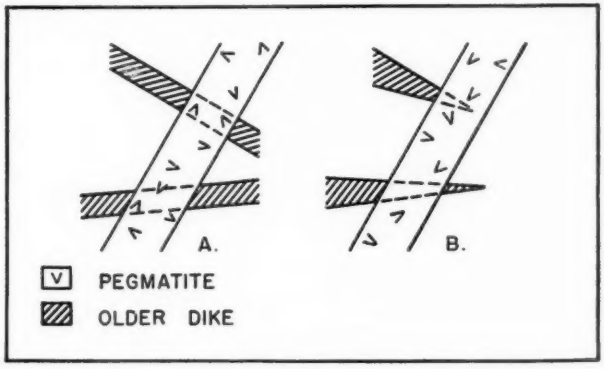


FIGURE 3.—EVIDENCE FOR NONDISPLACEMENT OF OLDER DIKES
(A) Two dikes having different attitudes are both cut by pegmatite. Considered stronger evidence for nondisplacement than Figure 1 *B*. (B) Upper dike missing on one wall. Lower dike narrower on one wall than on other. Dashed lines indicate portions of dikes replaced or incorporated rather than displaced.

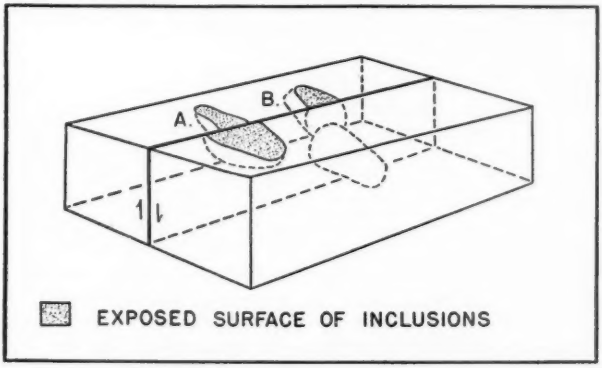


FIGURE 4.—FAULTING OF INCLUSIONS
Near-vertical fault cuts inclusions. Inclusion *A* is narrower on one side of fault, and inclusion *B* is "missing", having dropped below surface of exposure. Movement on fault shown by arrows. If walls of fault are later separated and pegmatite emplaced, resulting relations will suggest nondisplacement by analogy to Figure 3 *B*. (Modified from Wells and Bishop, 1954, Fig. 7).

eliminated that a tabular zone is absent on one wall owing to faulting of the missing part along the fissure horizontally beyond the limits of the exposure.

Matching of walls (*102*).—Descriptions of many narrow tabular bodies indicate that their walls match: removal of the body would allow the host rock to be fitted together perfectly.

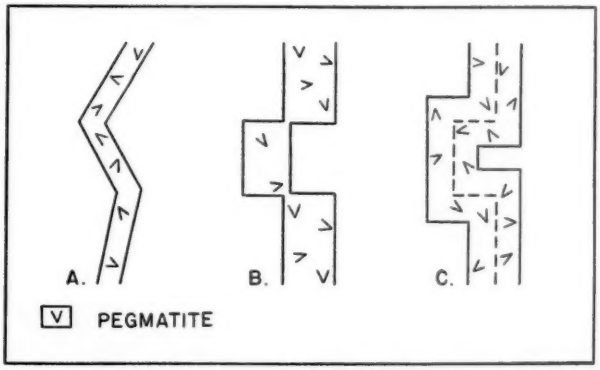


FIGURE 5.—PRINCIPLE OF MATCHING WALLS
(A) Fairly regularly shaped pegmatite. Uncertain whether walls separated or whether replacement outward from narrow crack occurred. (B) Irregularly shaped pegmatite. Walls definitely separated. Replacement outward evenly from crack would result in shape shown in (C), not (B).

Veins with matching, nonequidistant walls are considered by Hulin (1929, p. 38) and King (1948, p. 469) to denote displacement along irregular or undulating fractures. Pegmatites with matching walls which are very irregular (Fig. 5, *B*) are best explained as the effect of displacement (102); Figure 5, *A*, however, might result from stoping out of blocks or replacement spreading outward evenly in a straight line from a narrow crack.

Nonmatching walls do not prove lack of displacement. A pegmatite might force aside incompetent wall rocks irregularly. Displacement followed by stoping, assimilation, or metasomatism of parts of the wall rock would also result in nonmatching walls. For example, metasomatism spreading outward in straight fronts from an irregular crack shaped like that in Figure 5, *B* would produce Figure 5, *C*.

Relation of pegmatite and adjacent wall-rock composition (*112*).—Descriptions of some pegmatites note that chemical or mineralogical composition within the body, particularly near its border, varies sympathetically with the type or composition of the adjacent wall rock (112). As an illustration, the wall zone of the Kilton pegmatite, Grafton, New Hampshire, is richer in mica at places where micaceous schist rather than aplite is adjacent to the contact. Such relationships have been attributed to metasomatism (Jahns *in* Gilluly *et al.*, 1948, p. 94, referring to granites) and to assimilation of wall rock by a fluid pegmatite (Ball *in* Spurr and Garrey, 1908, p. 64). Sympathetic variations in pegmatite and wall-rock composition thus suggest nondisplacement—incorporation or replacement—of the host rock.

Absence of a relationship between pegmatite and adjacent wall-rock compositions does not distinguish displacement from nondisplacement, or mobility from nonmobility. Although Hanley (1951, p. 13) and Jahns and Wright (1951, p. 44) use the feature as evidence for lack of replacement, Hutchinson (1955, p. 20) states that pegmatites near Ross Lake, Northwest Territories, which include much evidence for replacement, usually are unchanged in composition where in contact with various wall-rock types. Moreover, Currier (1947, p. 80) and MacGregor and Wilson (1939, p. 208) suggest that metasomatism may produce a uniform rock composition in spite of original compositional variations. A uniform composition could result if equilibrium were established over the full volume of rock involved in the metasomatism, though different volumes of the different compositional types would have to be replaced.

General deformation of adjacent wall rocks and of pegmatite minerals (*104, 109*).—Fractured, bent, or closely folded wall rock is concentrated around the margins of many pegmatites and

modifies the regional structure. At least some displacement of the host rock is indicated, provided that the disturbance is not due to a post-emplacement deformation that crumpled incompetent wall rocks against a more competent pegmatitic mass (104).

Several features may establish whether post-pegmatite deformation occurred. Replacement or transection of deformed wall rock by pegmatitic material (Pl. 1, Fig. 2) has been thought difficult to explain by post-emplacement deformation (Stark and Barnes, 1932, p. 478; Chapman, 1941, p. 378). It should be reasonably certain that any replacing or transecting minerals crystallized essentially contemporaneously with the bulk of the pegmatite rather than later—for example, as a hydrothermal replacement.

Granulation or shearing of minerals late as well as early in the crystallization sequence of the emplaced body has suggested post-emplacement deformation (*e.g.*, Balk, 1937, p. 152). Conversely, the cementation of early, deformed minerals by later, undeformed minerals of the crystallization sequence has been attributed to deformation during crystallization (Maurice, 1940, p. 61; Page *et al.*, 1953, p. 19; Williams, Turner, and Gilbert, 1954, p. 25).

Absence of deformation within a pegmatite in spite of disturbance in the surrounding rocks implies that the wall-rock disturbance was not post-emplacement. However, caution must be used. Ramberg (1952, p. 256) believes that recrystallization during extreme deformation of a pegmatite can obliterate any drastic sign of cataclastic deformation in the resulting contorted body. He also suggests (1956, p. 195) that a pegmatite might not be conspicuously mylonitized if it had much less plasticity than the wall rock.

Lack of disturbance, other than post-emplacement deformation, in the wall rocks adjacent to a pegmatite might suggest nondisplacement of the host rock. Possibly, however, the rock moved outward in a large, undisturbed block, either expanded by regional tensional forces or wedged apart by the pegmatite. The volume increase would be adjusted by deformation at a considerable distance from the body. Balk (1937, p. 127) suggests that "long distance" effects of intrusion are exhibited around certain large intrusive bodies.

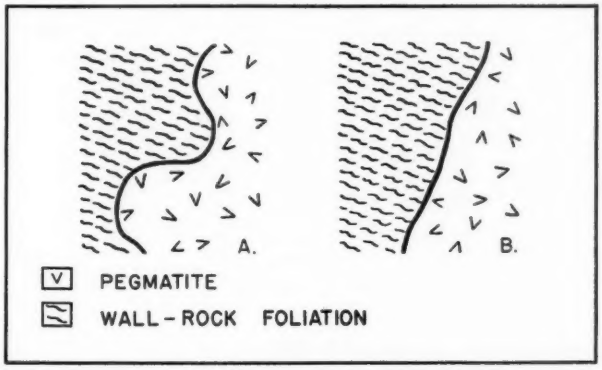


FIGURE 6.—TRANSECTION OF UNDISTURBED WALL-ROCK FOLIATION BY PEGMATITE
(A) Pegmatite contact irregular, nondisplacement indicated. (B) Contact straight, may result from nondisplacement mechanism or movement outward of large block of wall rock as fracture opened. Scale of drawing would be in feet. (*See* Fig. 17.)

If pegmatites contain irregular, rounded projections that transect essentially undisturbed wall-rock foliation (Figs. 6, *A*; 17) they are probably developed without displacement (109) (King, 1948, p. 460; Cameron *et al.*, 1954, p. 24). It is difficult to envision country rock expanded outward to form a very irregular opening or pushed outward by a very irregular body without the development of flexures or fractures along or across the foliation; however, a block might be displaced outward to make room for a reasonably straight tabular body without becoming deformed (Fig. 6, *B*; King, 1947, p. 63).

Not necessarily all the wall rock of Figure 6, *A* was displaced. Movement outward of host rock from a straight fracture, followed by engulfment of some wall rock by pegmatitic material invading irregularly outward beyond

the pre-existing opening, could produce the feature illustrated.

Conformance of wall rock to pegmatite contact (103).—At many pegmatites, the foliation of the wall rock diverges from its regional trend and bends to parallel the pegmatitic body, usually following most irregularities in the contact. Micaceous schists or other plastic, weak rocks are particularly likely to conform to pegmatite contacts (Pl. 1, Fig. 1). The foliation in places is contorted into small crinkles or drag folds. Unless caused by post-emplacement deformation, distorted rock molded around a pegmatite is evidence for displacement (103). If, in addition, the body is irregular in shape, various authors have attributed the feature to deformation by the emplaced body or to post-emplacement deformation.

Emplacement by injection of a fluid has been suggested by Jenks (1935, p. 181), Lahee (1941, p. 142), Jahns, Griffitts, and Heinrich (1952, Figs. 10, *B*, 24, *C*), and Noble (1952, p. 36). Ramberg (1952, p. 252) believes that the "force of crystallization" of solid, growing material may have pushed the walls outward. Foye (1922, p. 10) conceives that regional forces contemporaneous with emplacement may have acted upon both the still-fluid pegmatite and the wall rock and molded the foliation and the contact into parallelism.

Conversely, Greenly (1923, p. 348), Tolman (1931, p. 284), Grout *et al.* (1933), Nevin (1949, p. 196), and Ramberg (1955; 1956, p. 193) suggest that conformance of foliation to an emplaced body may in some cases result from post-emplacement deformation.

Criteria Pertaining to Forceful and Permissive Emplacement

Shape of pegmatite (201).—A pegmatite around which the wall-rock foliation is closely wrapped may have a very irregular shape, with many variously oriented, rounded convolutions along its margins (Fig. 7, *A*) (201). Not only displacement, but also forceful emplacement is denoted by this feature, provided that post-emplacement deformation cannot account for it. Stoll (1950, p. 4) believes that some irregular pegmatite shapes do not coincide with the shapes of openings likely to be produced in the wall rock by regional forces before emplacement. Force directed outward by an intruding or crystalloblastically growing pegmatite might produce such irregularities.

A regular lenslike or podlike shape, however (Fig. 7, *B*), can probably develop during either forceful or permissive emplacement, as suggested by Andersen (1931, p. 13). Regional compression acting parallel to the layers in a sequence of rocks of varying competency might buckle these layers into lens-shaped openings, provided that excessive external pressure acting perpendicular to the layering does not prevent this expansion. As an analogy, a stack of papers can be buckled to form openings between the sheets by compression parallel to the planes of these sheets.

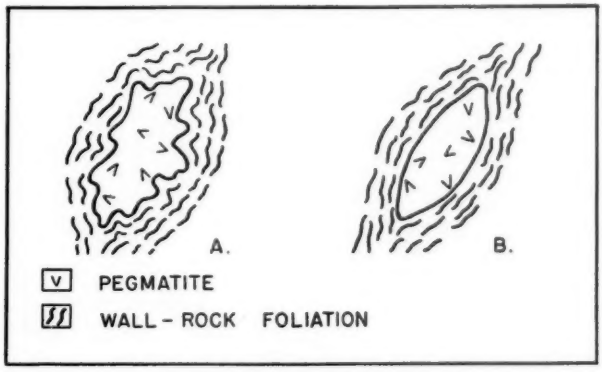


FIGURE 7.—SHAPE OF PEGMATITE

(A) Irregular, rounded shape, with wall-rock foliation deflected to follow contacts. Forceful emplacement indicated. (B) More regular, podlike shape which can probably be the result of either forceful or permissive emplacement. (It is assumed that the foliation in these two examples is not due to post-emplacement deformation.) Scale in feet.

Straight, tabular bodies have been interpreted variously as resulting from (1) metasomatic replacement spreading outward evenly from a fracture (Ramberg, 1952, Fig. 119) or controlled by a fissure zone (Webb, 1946, p. 184); (2) forceful wedging apart of a crack by pegmatitic fluid (Graton, 1906, p. 37; Lindgren, 1933, p. 152); (3) permissive draining of fluid into a fracture (Andersen, 1931; Emmons, 1940); (4) stoping of a large block (Billings, 1925, p. 149). Probably this feature is unsafe to use as a criterion for emplacement.

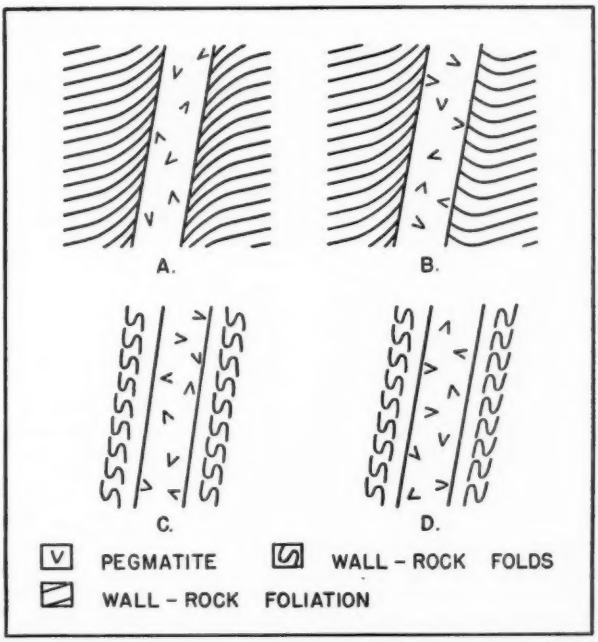


FIGURE 8.—BENDING AND DRAG FOLDING ALONG PEGMATITE CONTACT
(A) Direction of bending of foliation suggests faulting, does not distinguish forceful from permissive emplacement. (B) Direction of bending suggests forceful emplacement upward. (C) Drag folding suggests movement as in (A). (D) Drag folding suggests movement as in (B). Scale in feet.

Whether a pegmatite has a regular or irregular shape may depend on the amount of pegmatitic material emplaced as well as the structural control (Staatz and Trites, 1954).

Drag folds or bends adjacent to contact (202).—The pattern produced by drag folds or bends in the wall rock adjacent to a pegmatite may distinguish forceful from permissive emplacement. Foliation that is bent in the same direction on both walls of a pegmatite (Fig. 8, *B*) or drag folds that have a consistently opposite pattern on opposite sides (Fig. 8, *D*) indicate forceful emplacement provided they cannot be produced by post-emplacement deformation (202). The pegmatite, during intrusion, presumably dragged the adjacent rocks toward parallelism with its borders. (*See* Jahns, 1946, p. 215; Jahns, Griffitts, and Heinrich, 1952, Fig. 10, *A*; Page *et al.*, 1953, p. 16.)

Post-emplacement deformation might produce these patterns by shoving solid pegmatite through more plastic country rock or by causing shearing along one margin of a pegmatite which might reverse the direction of drag from a former pattern (Fig. 8, *A*, *C*). Post-emplacement deformation is unlikely where the dragged portions of wall rock are mineralized by pegmatitic material, or where pegmatite margins exhibit no evidence of shearing.

Bending of foliation in opposite directions on opposite walls (Fig. 8, *A*) could imply faulting either before or during emplacement (Chapman, 1941, p. 373; Jahns, Griffitts, and Heinrich, 1952, Fig. 10, *C*; Ramberg, 1952, p.

252; Griffitts and Olson, 1953, p. 211) or perhaps after emplacement. Similarly, an identical pattern of drag folds on both walls (Fig. 8, *C*) would not distinguish between forceful emplacement, permissive emplacement, or post-pegmatite deformation.

Thrusting in adjacent wall rock (203).—Thrusting of the adjacent wall rocks outward from or upward along the margins of a large intrusive suggests that the intrusive exerted considerable force on its surrounding rocks (Balk, 1937, p. 105; Waters and Krauskopf, 1941). Similarly, thrusting outward from pegmatites probably implies forceful as against permissive emplacement, particularly if the thrusting indicates that forces operated outward from both sides of the pegmatite (203). Evidence should indicate that the thrusting was not developed by a post-emplacement deformation.

Arrangement of pegmatites in regional fracture patterns (204, 205).—In some places pegmatites are arranged in patterns suggesting emplacement along openings primarily tensional in origin (204). Examples of such patterns are (1) cross joints in igneous host rock (Balk, 1937, p. 30), (2) cracks in igneous rock caused by unequal volume changes consequent on temperature variation (Andersen, 1931, p. 50), (3) echelon fractures (Cameron *et al.*, 1954, p. 24), and (4) openings between boudins (Ramberg, 1952, p. 252). Other types of openings might include (5) ring dikes, (6) fractures along the crests or troughs of folds that dip inward perpendicular to the planes of the folded beds, and (7) marginal joints along the borders of large intrusives.

Pegmatites developed along such fractures by a process involving displacement may be emplaced permissively rather than forcefully; regional tensional forces can create openings by dilation. However, pegmatites could enter narrow tensional cracks and force them apart, and the resulting body would be difficult to distinguish from one emplaced permissively. The criterion must be used with caution.

Pegmatites may follow regional shear patterns along which most movement was parallel to the walls of the fracture. If the pegmatite displaced its host rock, it may have developed forcefully or permissively depending on whether the shears or portions of a shear were tight and had to be wedged apart or whether they were open at the time of emplacement.

Permissive development may be postulated for bodies of pegmatite that fill interstitial spaces around disoriented breccia fragments (205); however, such fragments were possibly broken off the wall by the pegmatites during emplacement rather than formed during brecciation along pre-pegmatite faults or other openings. A study of the regional fracture pattern may help to determine the origin of the breccia, particularly if other faults or collapse structures barren of pegmatites occur in a recognizable pattern.

Criteria Pertaining to Mobility and Nonmobility

Inherited textures and structures (308).—The preservation within a pegmatite of textures, structures, or individual grain outlines originally present in the country rock (308) constitutes the strongest evidence for nonmobile emplacement. (*See* Cameron, Jahns, McNair, and Page, 1949, p. 89.) The inherited features may take the form of partial or complete pseudomorphs of individual minerals or of broad textural or structural patterns. Shainin (1948, p. 13) and Jahns, Griffitts, and Heinrich (1952, p. 9) cite examples. The relict features may also occur as networks of veinlets with nonmatching walls, in various stages of development in the host rock. Such networks have been considered satisfactory evidence for replacement (Lindgren, 1933, p. 184; Cameron, Jahns, McNair, and Page, 1949, p. 89).

Pattern of pegmatite foliation (301, 302, 309).—The pattern of foliation within a discordant pegmatite in relation to contacts and wall-rock structure may determine whether the material at the pegmatitic site was mobile during emplacement.

Foliation in an emplaced body that remains parallel to wall-rock structure across a discordant contact has been considered relict texture or an effect of post-emplacement deformation (Jahns, Griffitts, and Heinrich, 1952, p. 9). Foliation that does not extend through the entire body is probably relict texture. Nonmobility is indicated (309). Foliation should extend across the body if it is caused by post-emplacement deformation.

Foliation in an emplaced body that is consistently parallel to the contacts of the body though nonparallel to wall-rock structure has been attributed to flow during emplacement (Balk, 1937, p. 152; Jahns *in* Gilluly *et al.*, 1948, p. 93; Jahns, Griffitts, and Heinrich, 1952, p. 9). The feature denotes mobility (302).

The presence of appreciable primary-flow or deformation structure of any type, including primary foliation, protoclastic granulation, or fracturing, indicates mobility: at some time

during the emplacing process the material present flowed or was moved en masse (301).

Foliation in an emplaced body which is parallel to that of the wall rock and also to the contact along visible parts of the body may be relict texture, flow structure, or a consequence of post-emplacement deformation (Cloos, 1937, p. 46; Jahns, Griffitts, and Heinrich, 1952, p. 9).

Other methods of distinguishing between primary flow foliation, secondary foliation, and relict textures have been discussed by Trueman (1912), Balk (1937), Harker (1939), and Billings (1954).

Fractures within the pegmatite (304).—Fractures or fracture-filling bodies within a pegmatite which are structurally related to the contacts and linear features of the body (304) have been described by a few writers. These structures have suggested flowage of the pegmatitic material during crystallization (Page *et al.*, 1953, p. 19). For example, fractures that are consistently perpendicular to lineation in an emplaced body may be cross joints (Balk, 1937) which developed during flowage of the emplaced material. Other relationships of fractures to foliation and lineation might suggest longitudinal, marginal, diagonal, or primary flat-lying joints (Balk, 1937). If careful work establishes that the relationship is a primary flow pattern, mobile emplacement is indicated.

Orientation of inclusions (303, 310).—The foliation and lineation of inclusions in a nonfoliated pegmatite may be parallel to that of the wall rock (310). Many authors have attributed this relationship to metasomatic replacement or other nonmobile processes.

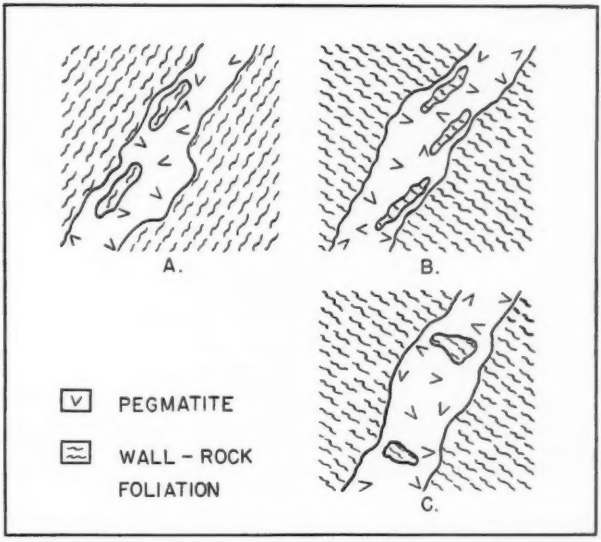


FIGURE 9.—ORIENTATION OF INCLUSIONS IN PEGMATITE

Foliation of inclusions is parallel to that of wall rock. (A) and (B) Long axes of inclusions parallel to direction of possible flow along pegmatite walls. Feature suggests nonmobility but could be produced by movement of inclusions. (C) Long axes not parallel to direction of possible flow. Nonmobility more definitely indicated. Scale in feet.

However, if the long axis or axes of these inclusions are parallel to a possible direction of flow in the pegmatite (parallel to a wall, for example), the inclusions may have moved along the pegmatitic channel without rotation (Figs. 9, *A*, *B*).

If the long axis or axes of inclusions are not parallel to a possible direction of flow in the pegmatite (Fig. 9, *C*) nonmobility is more strongly denoted.

Another qualification arises where the form and foliation of inclusions are parallel to the wall-rock foliation, but the lineation varies in orientation from one inclusion to another, though uniform in the country rock. Jahns (*in* Gilluly *et al.*, 1948, p. 93) believes that movement of the inclusions producing realignment of the foliation must have occurred; the line-

ation was rotated from its previous uniform orientation.

The foliation and lineation of some inclusions may be nonparallel to that of the wall rocks or other inclusions (303). This relationship suggests a mobile type of emplacement in which the inclusions were rotated from their original position. For example, Hutchinson (1955, p. 12) attributes differences in orientation of gneissosity of inclusions and of the wall rock to fluid emplacement. If, however, the near-by wall rocks were considerably distorted or brecciated before engulfment during emplacement, variations in foliation and lineation of the inclusions may merely reflect these disturbances. (*See* Perrin, 1956, p. 12; Ramberg, 1956, p. 200.) The feature should be used with caution.

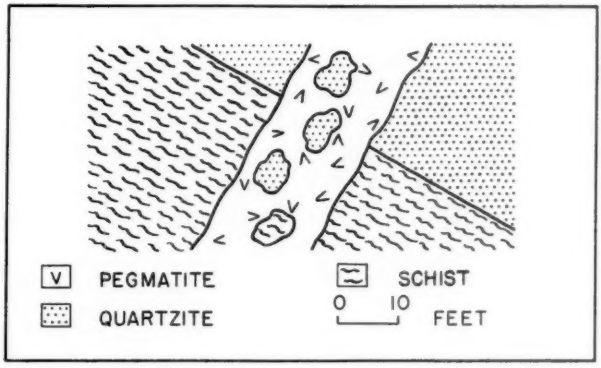


FIGURE 10.—COMPOSITION OF INCLUSIONS AND ADJACENT WALL ROCK
Some quartzite inclusions do not correlate with composition of adjacent wall rock and evidently were moved from original location. Feature suggests bodily mobility of pegmatitic material, particularly if large-scale.

Location of inclusions (*307*).—Inclusions within a pegmatite were presumably moved from their original positions if they are dissimilar in rock type or composition to any of the wall rock or to those regular units of the wall rock which, if extended into the pegmatite, would encompass these inclusions (Fig. 10). Mobile emplacement is inferred (307). It should be evident that the inclusions do not represent merely an altered phase of the host rock or inclusions present in the host rock before pegmatite emplacement (Ramberg, 1956, p. 200). Exposures should be good.

Internal zoning (*305*).—Many pegmatites consist of a series of mineralogically or texturally distinct shells or zones generally concentric around a core (305). This internal zoning has been discussed in detail by Cameron, Jahns, McNair, and Page (1949), who consider that it is a result of crystallization inward from the walls of a fluid-filled chamber. They cite the following features, combined, as supporting evidence: (1) the consistent sequence of mineral assemblages from the walls inward to the core; (2) the envelopment of the inner zones by the outer; (3) the transection of the outer zones by the inner but not the reverse; (4) the replacement of minerals of outer zones by those of inner zones but not the reverse; (5) the systematic variation in composition of plagioclase, beryl, and other minerals from the walls to the cores of some zoned pegmatites.

Ramberg (1956, p. 211) suggests that zones may form by migration of mobile ions into openings created by expanding joints, the zones crystallizing inward from the walls as the joint expands. To the extent that openings are present to attract the pegmatitic constituents, this mechanism would be classed as permissive emplacement.

The development of zones by crystallization of diffusing particles in expanding joints would seem a difficult explanation if the pegmatite is very irregular in shape, and if the wall-rock foliation deviates to remain parallel to the contact. Force from the pegmatite rather than expansion of the country rock by regional forces seems most capable of producing highly irregular, conformable bodies, provided the deviations in wall-rock foliation did not develop after emplacement (see discussion above on shape of pegmatite). The force involved is not

likely to be the force of crystallization of minerals growing outward in a solid medium, unless it can be satisfactorily shown that zones develop by crystallization outward. Crystallization from a pegmatitic fluid, as discussed below with reference to the Strickland pegmatite, would more reasonably explain this type of zoned body.

Crystallization into a fluid-filled or open chamber is a form of mobile emplacement of the material present; zoning is considered a criterion for mobility in distinguishing nondisplacement processes. Zoning may, of course, also develop during a process involving displacement.

If zones could develop as a result of a nonmobile process such as the migration inward of metasomatic "fronts", the transection of outer zones by inner zones, but not of inner by outer, is difficult to explain. In addition, a serious problem arises concerning the disposal of the constituents of the original rock, as material is being driven inward from all directions. If zones result from the migration of some ions inward and others outward, as suggested for some zones by Higazy (1949), it is puzzling that material from inner zones should replace or transect that in outer zones, without the reverse.

Only two concentric units, one of which is a narrow border zone, are present in some pegmatites. Because the border zone may reflect reaction with the wall rocks (Cameron, Jahns, McNair, and Page, 1949, p. 19; Murphy, 1949, M.S. Thesis, Dartmouth College; Jahns, Griffitts, and Heinrich, 1952, p. 23) the presence of only these two units in a pegmatite does not necessarily imply crystallization inward into a fluid-filled or open chamber.

Tapering pegmatite crystals (306).—Pegmatite crystals that are adjacent to the walls or to an outer unit of a pegmatite may be elongated perpendicular to the contact with that unit; these crystals may be tapered with the narrow ends abutting abruptly and evenly against the contact. The crystals widen inward from this contact and interfere with each other, and the inner edges terminate at varying distances from the contact (306). Figure 3 of Plate 1 illustrates the feature. These relationships have frequently been ascribed to the growth of crystals from the contact into a liquid or gas-filled space (Crosby and Fuller, 1897, p. 158; Bastin, 1911, p. 23–25; Stockwell, 1933, p. 33; Shaub, 1940; Jacobson and Webb, 1946, p. 33; Hanley, 1951, p. 13; Barth, 1952, p. 318; Griffitts and Olson, 1953, p. 221; Jahns, 1953, p. 584; Wells and Bishop, 1954, p. 104). Similar "comb structure" in hydrothermal veins has commonly been attributed to open space filling. (*See*, for example, Bateman, 1950, p. 108.) However, other authors have cautioned that "comb structure" in pegmatites may develop by replacement of the material into which the tapering crystals project (Jahns, Griffitts, and Heinrich, 1952, p. 44).

Thin sections of tapering crystals studied by the writer exhibit no evidence for replacement. Crystals remain normal to contacts even where contacts curve, and the orientation of the crystals has no consistent relationship to wall-rock foliation. Furthermore, in pegmatites examined by the writer, pegmatite minerals normal to and widening from a planar contact expand toward the center of a pegmatite (or toward the center of a fracture-filling body) rather than away from it. Crystal growth caused by replacement or metamorphic differentiation of pre-existing solid pegmatite or country rock presumably would not in all cases proceed into the pegmatitic mass: inward from a contact between pegmatite and wall rock, inward from a contact between two pegmatite units, and outward from inclusions.

Many pegmatites show tapered crystals in units that completely surround core units; this is difficult to explain by metasomatic replacement under open-system conditions (Stugard, 1955, U. S. Geol. Survey open-file report, p. 57).

Crystallization inward into a chamber filled with pegmatitic fluid or into an open cavity would logically explain all the relationships observed, and the feature is considered evidence for mobility in distinguishing nondisplacement mechanisms.

Criteria Pertaining to Introduction and Local Derivation

Bulk composition of pegmatite and wall rock (401, 403).—The bulk chemical composition of some types of host rocks, *e.g.*, some quartzites and marbles, is sufficiently dissimilar to that of enclosed granitic pegmatites to preclude the derivation of all the pegmatitic material from the enclosing rock, even by differential solution and migration of particles to the pegmatitic site. Such bodies have developed at least partly by introduction (401)—*See*, for example, Sederholm (1926, p. 137) on granites, and Engel and Engel (1953, p. 1080) on pegmatites and granites.

Some types of country rock, by differential solution and particle migration, could provide all the chemical constituents necessary for

development of enclosed granitic pegmatites. However, calculations may demonstrate that the ratio of pegmatitic material to total rock in the vicinity is too high to permit derivation of the entire pegmatitic mass from the rock originally present. Introduction of some material is required (403). The amount of introduced material would depend on the degree of difference between the composition of the pegmatites and that presumed to be representative of the original rock. Sederholm (1926, p. 137) states that in many migmatites that consist of a mixture of granite and leptite or metabasalt the granite makes up most of the rock. He infers that the granite was introduced. (*See* also Engel and Engel, 1953, p. 1080.) Careful analyses and a consideration of the relations in three dimensions are, of course, desirable.

Amount of pegmatitic constituents in adjacent wall rock (*402*, *404*).—Careful, systematic analyses may reveal that the wall rock adjacent to a pegmatite, compared to that in the general area farther from it, is impoverished in at least some of the chemical constituents, such as beryllium or boron, which are characteristically present in pegmatites (404). This relationship strongly implies the derivation of these pegmatitic constituents from the immediately surrounding rock. Greenly (1923, p. 345) and Ramberg (1952, Fig. 105) cite examples of pegmatites whose walls are impoverished in certain constituents, and they conclude that these bodies were derived locally.

Conversely, the adjacent wall rock may be enriched in pegmatitic constituents compared to the average country rock at some distance from the body (402). Aureoles of tourmaline-rich or beryl-rich wall rock adjacent to pegmatites have been noted and ascribed to introduction of constituents from sources other than the immediate walls of the bodies. (*See* Bannerman, 1943, p. 15; Cameron, Jahns, McNair, and Page, 1949, p. 11; Jahns, Griffitts, and Heinrich, 1952, p. 15; Page *et al.*, 1953, p. 17, 20.) Hutchinson (1955, p. 33) found that the concentration of beryllium in the wall rocks adjacent to certain pegmatites was greater than that farther from them, and he infers that the pegmatites have added beryllium to their walls. Many authors have described muscovitization and feldspathization of the wall rocks surrounding pegmatites.

Where evidence establishes that certain constituents, such as beryllium or boron, were emplaced during development of a pegmatite and are much more abundant in the immediately surrounding rock than in that farther from the body, introduction rather than local derivation is indicated. The introduced constituents were perhaps added to the wall rock from an intruding pegmatitic fluid, or possibly they were derived in the wall rocks at a considerable distance from the site of emplacement and migrated toward the site through a large volume of intervening country rock.

Criteria Pertaining to Host-Rock Breakdown and Residual Segregation

Physical state of wall rock before pegmatite emplacement (*501*, *502*, *503*, *504*, *505*).—Evidence that the rock surrounding a pegmatite was consolidated for a time immediately before emplacement of the body would prove that the pegmatite could not have crystallized as a locally derived residual segregation from a surrounding magmatic mass. Such evidence may include the presence of sediments or metasediments as wall rocks (501) or age determinations, either absolute or relative, which demonstrate that the pegmatite is appreciably younger than the surrounding rock (502). Comparative recency of the pegmatite would be established if, for example, a stage of intrusion or structural deformation definitely later than the major host rock rather than auto-injection or protoclastic with respect to it, occurred before emplacement of the pegmatite.

Evidence that the wall rocks were not solid throughout at the time of emplacement would denote residual segregation. Hypothetical cases might include transection or replacement of a pegmatite by material deemed to be late in the primary crystallization sequence of an enclosing host igneous rock (503); deflection or other modification of primary flow structure in host igneous wall rock where adjacent to the pegmatite contacts (504); or the occurrence of protoclastic textures and structures in a pegmatite that can be correlated with late-stage flow or other protoclastic structure in the host rock (505).

Application of Criteria to Pegmatites Studied

General Remarks

The writer attempted to apply the criteria discussed above to certain pegmatites in the field, in order to examine the practicality of the suggested approach and to determine as much as possible about the mechanisms of emplacement of specific pegmatites. The classification

of a pegmatite as to mode of emplacement will be affected by (1) degree of reliability assigned to each feature as a criterion, which is largely a matter of individual judgment, and (2) proportion of the pegmatite to which each feature can be applied. If a feature is present only at certain places within or near a pegmatite, it should denote a mode of emplacement only for the part of the body to which it relates.

The present investigation deals principally with features readily observable in the field. Undoubtedly other criteria could be added to the lists presented. For example, detailed chemical and mineralogical criteria might be used to shed additional light on the emplacement problem.

A number of pegmatites or closely related pegmatite groups were examined in Maine, New Hampshire, and Connecticut, principally during the summers of 1953 and 1954. Of these, 17 in New Hampshire and 6 in Connecticut (Figs. 12, 13) were chosen as best exposed and best illustrating the present approach to the emplacement problem. The conclusions drawn regarding modes of emplacement should be considered valid for those bodies only. Most of those studied are zoned or are closely related homogeneous bodies and are not necessarily typical of the majority of all pegmatites.

General features of the pegmatites are noted in Table 6. The general geology of all but three of these bodies was described and mapped during World War II studies (Cameron *et al.*, 1954); therefore, the present work could be concentrated on features relevant to modes of emplacement.

The three pegmatites not previously mapped are: a small body on the north side of Roaring Brook, .5 mile east of South Glastonbury, Connecticut; a pegmatite in a road cut on U. S. Highway 6A, .7 mile west of Cobalt, Connecticut; a small body on State Highway 10, 1.3 miles south of Gilsum, New Hampshire.

Regional Setting

All the pegmatites examined occur in highly folded and metamorphosed rocks: quartz-mica schists, granitic gneisses, and, less commonly, quartzites. The bodies studied in New Hampshire lie in and between the Keene and Grafton districts, along a north-northeastward-trending zone containing numerous pegmatites; those in Connecticut are located in the Middletown district. Figure 11 is an index map showing the districts studied.

The chief regional geologic features and the locations of the pegmatites studied are shown in Figures 12 and 13. Many articles on the geology of New Hampshire have been written by Billings and his co-workers. The reader is particularly referred to reports on Quadrangles in which the pegmatites investigated occur: Bellows Falls (Kruger, 1946a; 1946b); Sunapee (Chapman, 1952; 1953); Cardigan (Fowler-Lunn and Kingsley, 1937; Fowler-Billings and Page, 1942); and Rumney (Page, 1937, Ph.D. Thesis, Univ. Minn.; Fowler-Billings and Page, 1942). Articles on Connecticut include Rice and Foye (1927) and Foye (1949). A brief summary of the regional geology of the pegmatite districts of the two States appears in Cameron *et al.* (1954).

Articles on general features of the pegmatites of these areas include: Bastin (1910), Sterrett (1923), Chapman (1941), Olson (1942; 1950), Bannerman (1943), Cameron *et al.* (1945; 1954), Cameron and Shainin (1947), Murphy (1949, M.S. Thesis, Dartmouth College), and Stugard (1955, U. S. Geol. Survey open file report).

Pegmatites Illustrating Forceful Emplacement

General statement.—Five pegmatites for which forceful emplacement is considered to be an important, perhaps a dominant, mechanism are listed in Table 7, together with the features that indicate this mode of emplacement. These bodies are: the Bordonaro, Chandler Mills, Ruggles, and Strickland pegmatites, and the small pegmatite on Roaring Brook.

Strickland pegmatite.—The Strickland pegmatite (Fig. 14) serves as a good example of a body for which the criteria indicate forceful emplacement.

First to be examined is the displacement question, for which criteria from Table 1 are considered. The weight of the evidence observed favors displacement. The foliation of the schist surrounding the Strickland pegmatite is highly contorted adjacent to contacts in most places and has, in fact, become parallel to the contacts around most of the irregular body. This contortion is not the result of post-emplacement deformation because the distorted rock is replaced by tourmaline, muscovite, and feldspar, the crystallization of which involved emanations introduced at the time of emplacement. For example, many large, undeformed black tourmaline crystals transect the structure of the deformed schist. Much of the tourmaline in and around the pegmatite evidently developed during the primary crystallization of

TABLE 6.—GENERAL FEATURES OF PEGMATITES STUDIED

Pegmatite*	State	Quadrangle	Wall rock	Approximate orientation of pegmatite†	Approximate orientation of wall rock
Bordonaro	Conn.	Middle Haddam	Schist, gneiss	N. mod. W.	N. mod. W.
Britton	N. H.	Bellows Falls	Schist, gneiss, minor quartzite	NE. mod. NW.	NW. mod. NE. to mod. SW.
Case No. 1	Conn.	Middle Haddam	Gneiss	N. mod. E.	NE. gen. NW.
Chandler Mills	N. H.	Sunapee	Schist	var. NE. gen. SE.	var. NE. mod. SE.
Coleman	N. H.	Cardigan	Schist, granite	N. vert.	NE. st. NW.
Danbury	N. H.	Cardigan	Gneiss	N.–NE. mod.–st. W.	N.–NE. mod.–st. E.
E. E. Smith	N. H.	Cardigan	Gneiss, schist	NE. st. NW.	NE.–NW. mod.–st. E.
Gillette	Conn.	Haddam	Schist, quartzite, gneiss, marble	NW. vert.	NW. st. NE.
Pegmatite on Highway 6A	Conn.	Middle Haddam	Gneiss	NE. st. NW.	NW. st. NE.
Pegmatite on Highway 10	N. H.	Bellows Falls	Schist	NE. mod. SE.	NW. mod. NE.
Hoyt Hill	N. H.	Cardigan	Schist, granite	N. vert.	N. st. W.
Kilton	N. H.	Cardigan	Schist, aplite	NE. mod.–st. SE.	NE. st. SE.
Kirk No. 1	N. H.	Bellows Falls	Schist	NW. mod. NE.	NW. mod. NE.
Kirk No. 2	N. H.	Bellows Falls	Gneiss	NE. st. SE.	N.–NW. mod.–st. E.
Marston	N. H.	Cardigan	Gneiss, minor aplite	NW. mod. NE.	NE. mod. SE.
Nancy No. 2	N. H.	Rumney	Schist	NE. st. NW.	NE. st. SE.
Palermo No. 1	N. H.	Rumney	Schist	NE. mod. SE.	NE. st. SE. to st. NW.
Paul	N. H.	Sunapee	Gneiss	NE. st. NW.	NE. mod. SE.
Rice-Palermo No. 26	N. H.	Cardigan	Schist	N. st. E.	N. st. E.
Pegmatite on Roaring Brook	Conn.	Glastonbury	Gneiss	NE. st. NW.	NE. mod. NW.
Ruggles	N. H.	Cardigan	Schist	NE. var. SE.	NE. st. SE.
Strickland	Conn.	Middle Haddam	Schist, pegmatite	NW. mod.–st. SW.	NW. mod.–st. SW.
Wadhams-Tucker	N. H.	Cardigan	Gneiss	E. mod. N.	NW. var. SW.

* If two or more pegmatites are present at a given location, that referred to is the one which contains the chief mine or prospect. The names "Marston" and "E. E. Smith" are exceptions; they refer to groups of three pegmatites each. Consult Cameron *et al.* (1954) for fuller descriptions.

† Direction of strike is listed first, followed by direction of dip. Abbreviations are as follows: gen. = gentle, mod. = moderate, st. = steep, var. = variable.

the body rather than later. Jenks (1935) cites the consistent decrease in refractive indices and birefringence of tourmaline in the Strickland pegmatite with time as determined from the paragenetic sequence. He believes that much of the black tourmaline crystallized during a magmatic phase and that the tourmaline in the

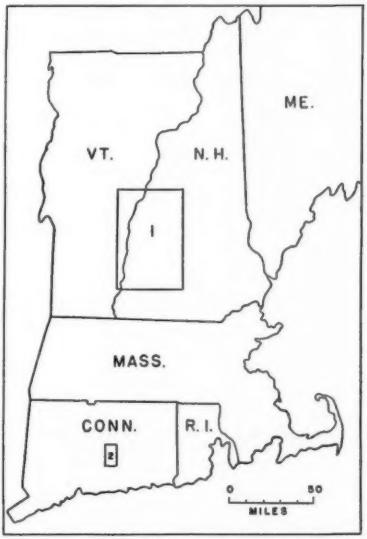


FIGURE 11.—INDEX MAP OF PART OF NEW ENGLAND
Area encompassed by number "1" is detailed in Figure 12; number "2" indicates Figure 13.

schist developed early in the period of intrusion. The tourmaline in the schist is also black and has the same refractive indices and birefringence as tourmaline in the pegmatite which is early in the paragenetic sequence; tourmaline late in the paragenetic sequence is blue and has lower refractive indices and birefringence.

Another line of evidence in conflict with post-emplacement deformation is the general lack of deformation of the pegmatite minerals.

Displacement thus appears significant in the development of the Strickland pegmatite, though the possibility remains that some country rock was incorporated into the body rather than displaced. Whether the entire mass of country rock originally occupying the pegmatitic site can be accounted for in the present volume of distorted wall rock is uncertain from available exposures. This same qualification must be applied to the other pegmatites studied where wall-rock foliation is deflected to parallel pegmatite contacts.

The irregular, embayed outline of the Strickland pegmatite denotes forceful as opposed to permissive emplacement. An opening of such irregularity would probably not form in the wall rock before the pegmatite came into place. No features favoring permissive emplacement were observed.

A consideration of the criteria from Tables 1 and 2 has thus led to the conclusion that forceful emplacement is a significant, perhaps the major, mechanism by which the Strickland pegmatite came to occupy its present site. Forceful emplacement may mean (1) that the pegmatitic material, crystallizing either from a fluid or in an essentially solid medium, was able to force aside the host rock, or (2) that regional forces contemporaneous with emplacement deformed the pegmatite and the schist simultaneously.

At this pegmatite, further conclusions may be drawn concerning the type of forceful emplacement. It is suggested that the observed relations are more reasonably explained if the Strickland pegmatite crystallized from a fluid rather than in a solid medium. This is because the body contains undeformed, generally concentric zones.

If the pegmatite grew outward from a center in the solid state, pushing aside the country rock by a "force of crystallization", evidence should demonstrate that the inner zones crystallized before the outer ones. The zoning indicates the opposite relationship.

It may be assumed that crystallization in the solid state proceeded inward from the pegmatite walls. If the force of crystallization produced the intense deformation of the adjacent wall rocks, then at the time the inner zones developed, the crystallizing minerals in these zones should have exerted force on the already crystallized outer zones and caused them to be strained, bent, or fractured together with the wall rocks.

Similarly, if regional forces contemporaneous with emplacement deformed the host rock and the solid, growing pegmatitic body simultaneously, the pegmatite grains should be strained or fractured, and the zones, particularly the outer ones, should be brecciated or otherwise deformed. The outer zones should not have remained parallel to the walls of the chamber while it became increasingly irregular.

Such deformation did not occur at the Strickland pegmatite. The pegmatitic grains are in general undeformed, and the zones, especially the outer ones, closely parallel the irregular contacts of the body. These relations strongly suggest that the forces causing deformation of the wall rock were effective before

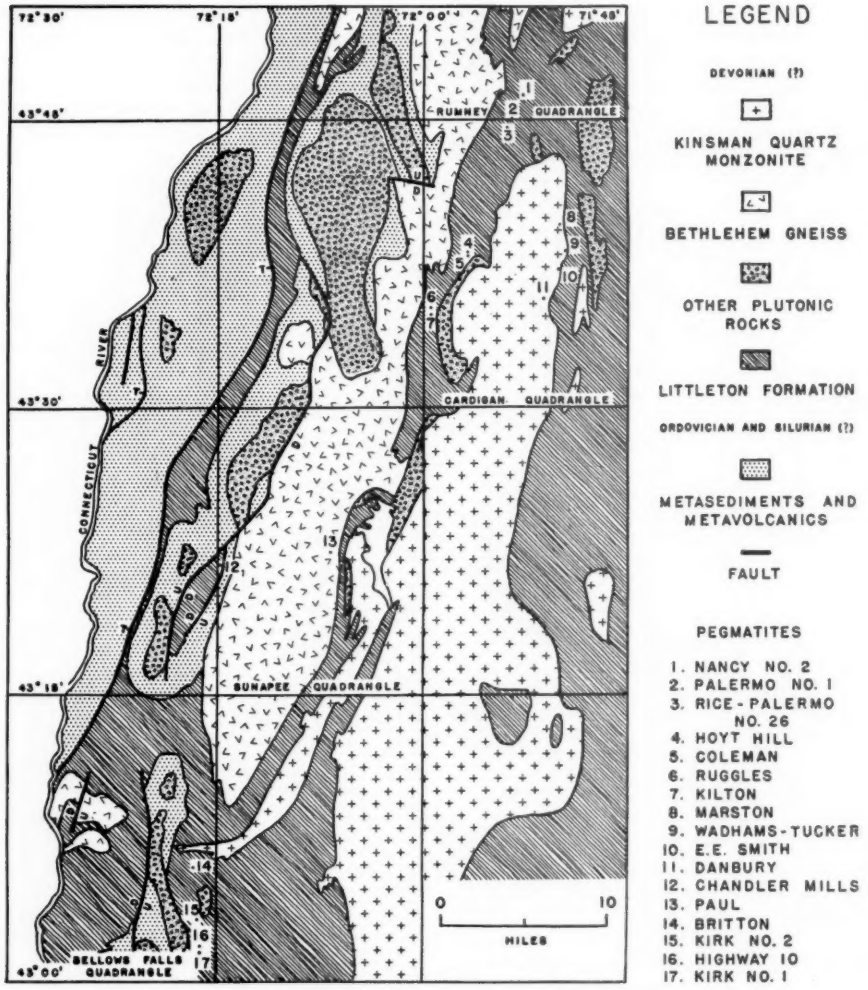


FIGURE 12.—GEOLOGIC MAP OF PEGMATITE DISTRICTS IN WEST-CENTRAL NEW HAMPSHIRE

Location of pegmatites studied are shown by numbers. Corresponding pegmatite names are in legend. (Modified from Billings, 1955.)

any crystallization—that a pegmatitic fluid filled the chamber and either bulged out the host rock or was deformed with it while still in a fluid state. After the irregularly shaped chamber was produced, the zones proceeded to crystallize in essentially undisturbed, concentric units closely conformable to the irregular walls of the chamber.

Ramberg (1956) states,

"Crumpling and plastic distortion of schists and gneisses adjacent to pegmatite bodies . . . have been taken as indications of forcefully injected pegmatite magma. It seems, however, unbelievable that a fluid melt could cause intense plastic deformation in solid rocks. (p. 198)

"It is physically impossible that the kind of pinch-and-swell structure shown in Plate 3 could develop by injection of the pegmatite as a melt." (p. 194)

However, evidence is offered by laccoliths, for example, that fluids have been able to arch solid host rocks. One might also point to experiments in which minerals and rocks are compressed by subjecting them to hydrostatic pressure transmitted through a fluid (*e.g.*, Birch *et al.*, 1942). The writer sees no reason why a pegmatitic fluid, perhaps but not necessarily magmatic, could not have bent aside and locally crumpled incompetent schistose host rock to create a chamber from which crystallized the Strickland pegmatite.

In addition to forceful emplacement, minor nondisplacement of the nonmobile type is demonstrated by the occurrence of relict textures locally within the Strickland body near the contacts with the surrounding rock and with inclusions.

Other pegmatites.—The evidence in favor of forceful emplacement is in general similar at the Bordonaro, Chandler Mills, and Ruggles pegmatites (Table 7). All three bodies have irregular walls along which the structure of the country rocks has been modified. The transection of these modified structures by undeformed tourmaline crystals demonstrates that post-emplacement regional forces did not produce the modifications. However, the local transection of essentially undisturbed wall-rock foliation by irregular contacts and the presence

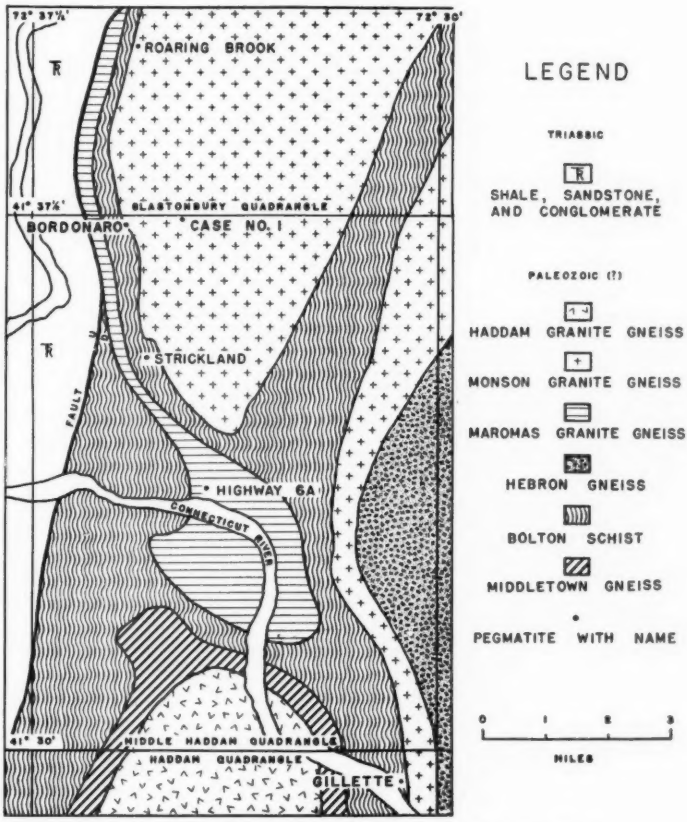


FIGURE 13.—GEOLOGIC MAP OF PEGMATITE DISTRICT NEAR MIDDLETOWN, CONNECTICUT

Locations of pegmatites studied are shown on map. (Slightly modified from Cameron *et al.*, 1954, Fig. 6.)

TABLE 7.—CRITERIA FOR EMPLACEMENT PRESENT AT PEGMATITES STUDIED

	Bordonaro	Chandler Mills	Roaring Brook	Ruggles	Strickland	Kirk No. 1	Rice-Palermo No. 26	Danbury	Gillette	Wadhams-Tucker	Britton	Case No. 1	E. E. Smith	Kilton	Marston	Nancy No. 2	Palermo No. 1	Coleman	Hoyt Hill	Kirk No. 2	Paul	Highway 6A	Highway 10
(1A) Displacement																							
101. No wall-rock units missing	•	•	•	•	•	•	×	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
103. Wall-rock foliation deflected to parallel contact	*×	×	○	×	×	×	×	○	○	•	•	•	•	○	•	•	•	•	○	•	•	•	•
104. Localized disturbance of wall rock	*×	×	×	×	×	×	×	×	×	×	○	○	○	○	○	○	○	○	○	○	○	•	•
(1B) Nondisplacement																							
107. Wall-rock unit missing, pegmatite occupies space	†×	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
108. Evidence of nonmobile emplacement	○	○	•	○	○	○	○	○	○	○	•	○	○	○	○	○	○	○	○	○	•	×	×
109. Irregular contact cuts undisturbed wall-rock foliation	•	○	•	○	•	•	•	×	×	×	•	•	×	×	○	•	•	•	○	•	•	×	×
112. Pegmatite, wall-rock composition related	•	•	•	•	•	•	•	•	•	•	•	•	•	×	•	•	•	•	•	•	•	•	•
(2A) Forceful																							
201. Pegmatite shaped irregularly	*×	×	•	×	×	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
202. Bends or drag folds suggest intrusion from one direction	•	•	×	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
203. Wall rock thrust outward from pegmatite	•	•	○	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•
(2B) Permissive																							
204. Pegmatite follows regional tension fracture pattern	•	•	•	•	•	•	•	•	•	•	•	•	×	•	×	•	•	•	•	•	×	•	•

(3A) Mobility																							
303. Wall rock, inclusion structures not parallel	•	•	•	•	•	•	•	•	○	•	•	•	•	•	•	•	•	•	•	•	•	•	×
305. Internal zoning	×	×	•	×	×	×	×	×	×	×	×	×	×	×	×	×	×	•	•	•	•	×	×
306. Pegmatite minerals widen inward	×	○	•	○	×	○	×	×	○	×	×	×	×	×	×	×	○	×	○	○	×	○	•	•
(3B) Nonmobility																							
308. Relict textures or structures	○	○	•	○	○	○	○	○	○	○	•	○	○	○	○	○	○	○	○	○	•	×	×
309. Wall rock, discordant pegmatite foliation parallel	•	•	•	•	•	•	•	•	•	•	•	•	•	•	•	○	•	•	•	•	•	×	•
310. Wall rock, inclusion structures parallel	•	•	•	•	•	•	•	•	○	•	•	•	•	•	•	○	•	•	•	•	•	×	×
(4A) Introduction																							
402. Adjacent wall rocks enriched in pegmatitic constituents	×	×	•	×	×	×	×	×	×	×	○	×	×	×	○	×	×	×	×	×	•	×	×
Dominant modes of emplacement	(2A)					(2A) or (2B)		(4A) and either (2A) or (2B)			(4A), (2A), or (2B)							(1A) or (1B)				(1B)	

× = feature widespread at pegmatite
○ = feature very local only at pegmatite
• = feature not observed at pegmatite
* = refers to schist only
† = refers to gneiss only

locally of relict textures suggest that small parts of the bodies developed without displacement of the host rocks.

The Bordonaro pegmatite (Fig. 15) transects a contact between relatively incompetent quartz-mica schist and more competent biotite granite gneiss. Where the host rock is schist, the pegmatite is very irregular, and the schist foliation in many places conforms to the contact. Pegmatitic constituents have mineralized the deformed schist. Forceful emplacement is indicated. However, a straight, narrow portion of the pegmatite extends into the gneiss parallel to the gneissosity. The gneissosity appears to strike into the northern exposed nose of the pegmatite without deflection; this suggests lack of displacement of the gneiss. Zoning in this portion of the body demonstrates that its emplacement was mobile, and the enrichment in tourmaline of the gneiss near the contact indicates that at least some introduction of material occurred. The Bordonaro pegmatite

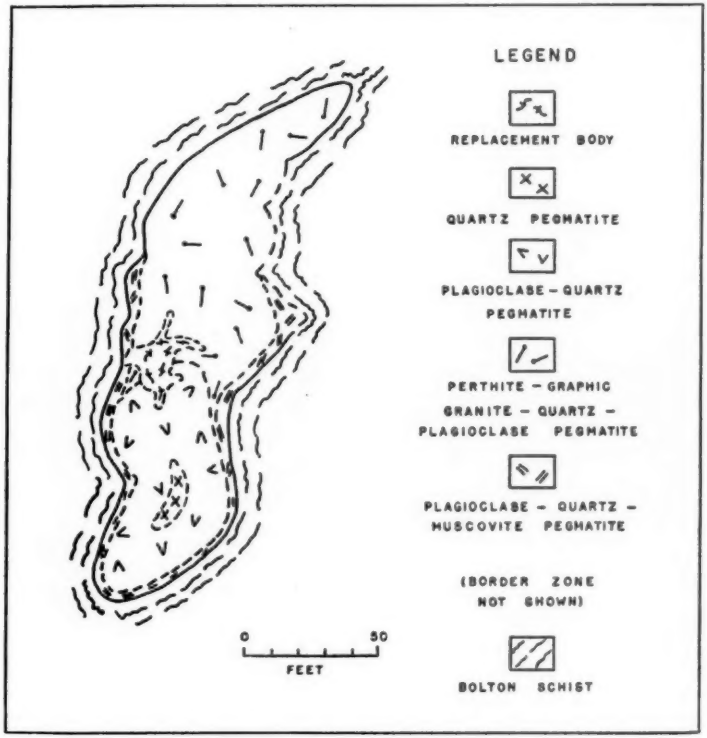


FIGURE 14.—STRICKLAND PEGMATITE

Composite cross section of Strickland pegmatite. Note generally concentric nature of zones (replacement body excluded) and conformance of schist foliation to irregularities of pegmatite boundary. Border zone, too thin to be shown on map, is closely conformable to pegmatite contact and just inside it. (From Cameron, Jahns, McNair, and Page, 1949, Fig. 29.)

PLATE 1.—FEATURES ILLUSTRATING MODES OF EMPLACEMENT

FIGURE 1.—Foliation of schist is deflected to parallel contact of pegmatite (white). More massive wall-rock layers are quartzite. Ruggles pegmatite, Grafton, New Hampshire.

FIGURE 2.—Locally contorted schist (upper left) is replaced by undeformed tourmaline crystals (black, center of picture) along schist-pegmatite contact. Feature indicates that deformation of schist was not post-emplacement. Chandler Mills pegmatite, near Newport, New Hampshire.

FIGURE 3.—Sharp contact at Coleman pegmatite, Orange, New Hampshire. Pegmatite minerals are oriented perpendicular to contact and widen from them into the pegmatite (downward in photograph). Wall rock is granite. Feature indicates mobility.

FIGURE 1

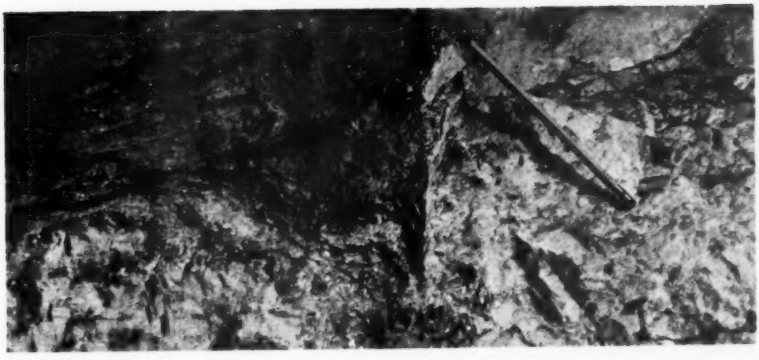


FIGURE 2

FIGURE 3

FEATURES ILLUSTRATING MODES OF EMPLACEMENT

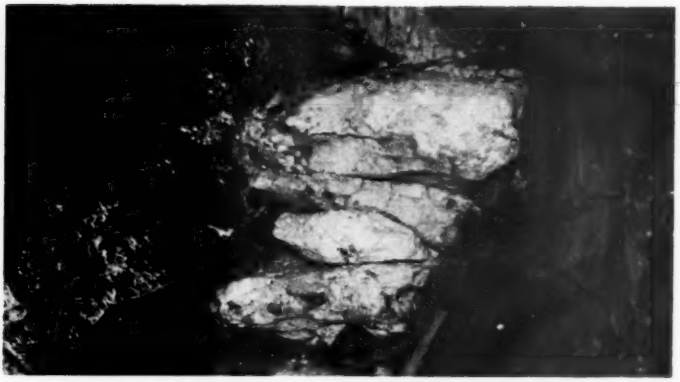


FIGURE 1

FIGURE 2

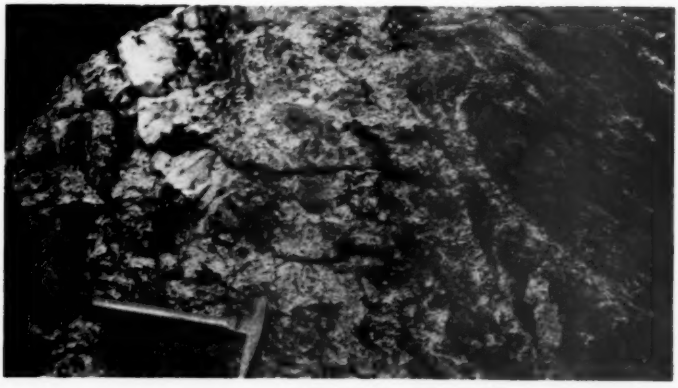


FIGURE 3

FEATURES ILLUSTRATING MODES OF EMPLACEMENT

thus seemingly forced aside incompetent schist during emplacement but did not deflect the structure of the more competent gneiss.

At the pegmatite on Roaring Brook, the gneiss foliation is deflected in the same direction

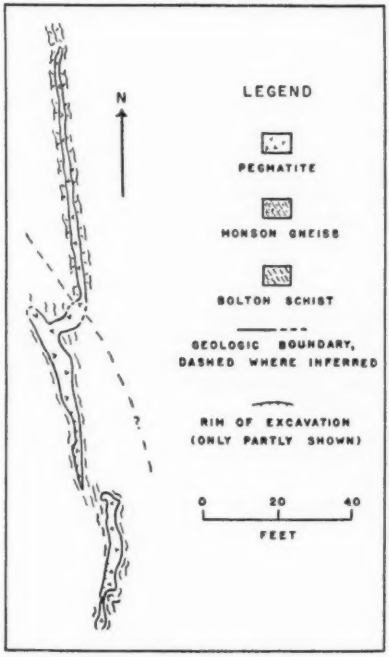


FIGURE 15.—BORDONARO PEGMATITE
Note difference in shape of pegmatite and in degree of bending of wall rock between portions of pegmatite in gneiss and in schist. (Modified from Cameron *et al.*, 1954, Fig. 128.)

along the contacts on opposite walls of the body. This bending is abrupt on one side, though broader on the other, and it persists along the entire exposed contact (Fig. 16). It is presumably not a result of post-emplacement deformation because of lack of deformation in the pegmatite itself and lack of shearing along the contacts. At several places there is evidence of small-scale thrusting outward and upward from the western contact. The pegmatite evidently moved downward during at least the latter stages of its emplacement, dragging the adjacent wall rock downward. It probably expanded locally during emplacement, wedging small blocks of wall rock outward and upward. The evidence implies forceful emplacement.

Pegmatites Illustrating Forceful or Permissive Emplacement

At two of the pegmatites studied, the Kirk No. 1 and the Rice-Palermo No. 26, features from Table 1 indicate that displacement of the host rock is the most significant process. At the Rice-Palermo No. 26, schist layers split around the northern exposed nose of the body; at the Kirk No. 1 the schist is locally contorted along the contact, and the contorted rock is transected by material that evidently emanated from the pegmatite. However, a choice between forceful and permissive emplacement could not be made owing to lack of distinguishing criteria. As in the preceding group, local features at these pegmatites suggest that minor portions of the bodies, particularly near the contacts, were developed without displacement of the country rock. Critical features observed at these bodies are listed in Table 7.

Pegmatites Illustrating Forceful or Permissive Emplacement and Also Mobile Introduction

Some features present at three pegmatites, the Danbury, Gillette, and Wadhams-Tucker (Table 7), favor displacement of the country rock. However, appreciable portions of the bodies appear to be the result of introduction.

Displacement is indicated by deformed wall rock near the contacts which is replaced by pegmatitic material. The amount of structural disturbance varies from place to place with the competence of the wall rocks. Where the wall rock is incompetent micaceous schist, its foliation near the pegmatite contact is in many places bent toward conformity with the outlines

PLATE 2.—FEATURES ILLUSTRATING MODES OF EMPLACEMENT

FIGURE 1.—Schistose wall rock bends around irregularities in Gillette pegmatite, East Hampton, Connecticut, as at top of photograph. Compare with Figure 2 of Plate 2.

FIGURE 2.—Near-vertical foliation of gneissose wall rock abruptly and somewhat irregularly transected by Gillette pegmatite. Pegmatite has replaced gneiss locally outward from contact along foliation planes. Figures 1 and 2 of Plate 2 illustrate typical contrast in behavior of incompetent and competent wall rock at many pegmatite contacts.

FIGURE 3.—Small pegmatite on Highway 10 near Gilsum, New Hampshire. Relict textures of wall rock and parallelism of wall-rock foliation to that of inclusions (center and right side of picture) suggest nonmobility for outer portion of body. However, small white quartz pod or "core" in pegmatite (upper left of photograph) suggests mobility for central part of body. Figure 18 is sketch of this pegmatite.

of the body. Where schist is interbanded with more competent gneiss, the foliation is more gently warped. For example, the foliation at the Wadhams-Tucker pegmatite changes gradually from a steep southward or almost vertical dip 50 feet north of the body to a gentle southward and in places horizontal to gentle northward dip near the northward-dipping hanging-wall contact. Similar bending occurs along portions of the Danbury pegmatite to a distance of 15 feet outward from the contact. The gentle flexing at these bodies strongly suggests at least partial displacement of host rock.

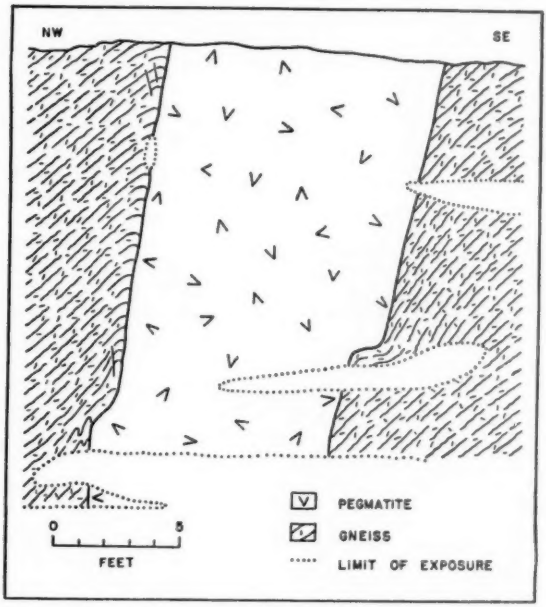


FIGURE 16.—PEGMATITE ON ROARING BROOK
Vertical section. Bending of wall-rock foliation suggests forceful emplacement in downward direction.

However, where relatively competent gneisses are the dominant wall-rock type, rounded projections of the pegmatites may transect the gneissic foliation; the latter is not noticeably flexed or sheared. The evidence favors incorporation of appreciable quantities of competent country rock into these pegmatites. It is difficult to conceive of a mechanism whereby a large volume of country rock is moved outward by an irregular pegmatite or to form an irregular opening without being bent or fractured. The sparseness of inclusions suggests that much of the host rock was digested.

The relations noted at the Gillette, Danbury, Wadhams-Tucker, and Bordonaro pegmatites are also present at some of the other bodies studied. Schistose wall rocks in many places are bent toward parallelism with the pegmatite contacts, whereas more competent rocks such as quartz-feldspar gneisses or quartzites are generally abruptly transected by portions, including rounded bulges, of the bodies. (Compare Figs. 14 and 17; Pl. 2, figs. 1 and 2.) The relations observed suggest that units of weaker wall rocks were moved aside to make room for pegmatites, whereas the more competent rocks were not so easily displaced and were in many cases incorporated into them.

The Danbury, Gillette, and Wadhams-Tucker bodies evidently developed to an appreciable degree without displacement of the host rock; criteria from Tables 3 and 4 help to determine nondisplacement mechanisms. Zoning and the widening inward from the contacts of pegmatite minerals in many places indicate mobility during emplacement. Macroscopically visible tourmalinization, feldspathization, and muscovitization are present in the wall rocks at various places along the contacts. Introduction

rather than local derivation evidently took place, and criteria from Table 5 need not be considered.

Pegmatites Illustrating Forceful or Permissive Emplacement or Mobile Introduction

At seven of the pegmatites studied, evidence concerning displacement is ambiguous, but several features denote mobility during the emplacement (Table 7). These pegmatites are: the Britton, Case No. 1, E. E. Smith, Kilton, Marston, Nancy No. 2, and Palermo No. 1.

Flexing and fracturing of wall rock near the contacts of these bodies, though locally present, are generally lacking. At most of the bodies, the contorted wall rocks are replaced by abundant tourmaline and feldspar presumed to represent additions from the pegmatites. The deformation was thus not post-emplacement. Alternate interpretations of the sparseness of contortions are possible: the pegmatites did not displace country rock, on the whole; or the bodies, where not too irregular, wedged aside their host rock or filled pre-existing openings.

At the E. E. Smith and Kilton pegmatites, considerable host rock was probably never displaced. Rounded, irregular projections of these bodies transect the wall-rock foliation without appreciably bending or fracturing it in many places (Fig. 17). However, in other spots small irregularities in the contacts are angular and seem fracture-controlled; displacement may have occurred at such locations.

Large muscovite crystals in the border and hanging-wall mica zones of portions of the Kilton pegmatite are more abundant where the body contacts schist rather than where aplite is the wall rock. Local incorporation as opposed to displacement of the country rock by the pegmatite is suggested. Much of this muscovite is in zoned portions of the pegmatite, and many of the crystals are oriented perpendicular to the contact, interfering with each other as they widen inward from it. The wall-rock foliation is in many places parallel to the contact and has no relation to the orientation of the muscovite crystals. The features favor growth inward into a fluid-filled chamber developed in part by incorporation of wall rock, rather than metasomatic growth of pegmatite crystals in host rock.

To the extent that displacement occurred, the E. E. Smith and Marston bodies were perhaps developed permissively rather than forcefully because of their echelon arrangement, which suggests tension fracture control (Cameron *et al.*, 1954, p. 24). Possibly, however, they wedged apart initially narrow cracks. At the other bodies, no features pertaining to forceful or permissive emplacement were observed.

All the pegmatites of the group are zoned and contain crystals that widen inward from the wall-rock contacts in many places. Mobility was evidently the dominant process not involving displacement. An outer aplite unit of the Nancy No. 2 pegmatite has in many places a relict texture. Possibly much host rock at the site of the present outer portion of the body was never mobile, though the major part of the pegmatite is well zoned.

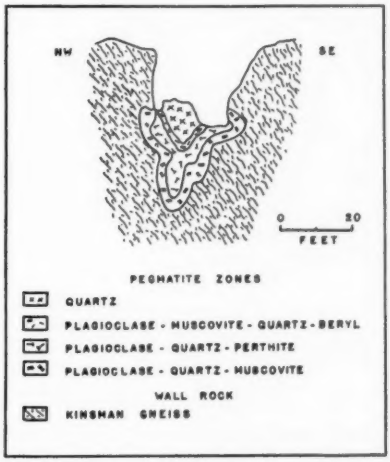


FIGURE 17.—E. E. SMITH PEGMATITE
Vertical section of northernmost of the three E. E. Smith pegmatites. Note well-developed zoning and abrupt transection of gneiss foliation by rounded irregularities of pegmatite. Features suggest a nondisplacement mechanism of emplacement. (From E. N. Cameron, unpublished map)

One or more of the minerals tourmaline, beryl, feldspar, and muscovite are concentrated in the wall rocks adjacent to all the pegmatites in the group; this strongly suggests that introduction was a significant mobile process. The lack of inclusions, particularly in pegmatites that have exposed floors (E. E. Smith) or that narrow rapidly downward (Nancy No. 2), suggests that stoping without assimilation was not a major process.

In summary, the seven pegmatites in this group have evidently formed in large part by forceful emplacement, permissive emplacement, or mobile introduction, or by some combination of these mechanisms.

Pegmatites Emplaced With or Without Displacement of Host Rock

Four pegmatites, the Coleman, Hoyt Hill, Kirk No. 2, and Paul, have few diagnostic features (Table 7). The bodies may or may not have displaced significant portions of their host rocks, and the degree of mobility is also problematical.

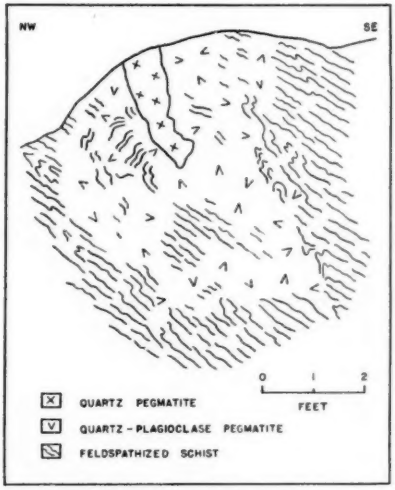


FIGURE 18.—PEGMATITE ON HIGHWAY 10
Vertical section. Relict textures in outer part of pegmatite suggest nonmobility for that portion, but quartz pod suggests mobility for central part of body.

Minor bending of wall-rock foliation, evidently not post-emplacement, occurs locally along the contacts of the pegmatites; this suggests some displacement of wall rock. Most of the wall rock appears undisturbed. Criteria distinguishing forceful from permissive emplacement are generally lacking, though the Paul pegmatite is in echelon arrangement with its associated bodies. The pattern suggests permissive emplacement into pre-existing tensional openings to the extent that the host rocks were displaced.

The amount of mobility that developed during emplacement is uncertain; none of the bodies can be considered well zoned, yet none contains appreciable amounts of relict texture from the country rock. At each body, pegmatite minerals project and widen inward from the contacts locally, but in most spots such orientation is lacking. The Coleman, Hoyt Hill, and Kirk No. 2 pegmatites were evidently not locally derived, as the adjacent walls are enriched in pegmatitic constituents. The evidence for introduction is less conclusive at the Paul pegmatite, though rocks adjacent to some of the near-by pegmatites of the echelon group are tourmalinized and muscovitized.

Pegmatites Emplaced Without Appreciable Displacement of Host Rock

At two pegmatites, on Highway 10 in New Hampshire and Highway 6A in Connecticut, evidence favors lack of displacement of the country rock in large part (Table 7). Pegmatite contacts transect undisturbed wall rock irregularly. Relict textures are abundant in the outer portions of the bodies, though evidence suggests that the central parts developed by mobile means.

At the Highway 10 outcrop, veinlets and irregular patches of quartz, plagioclase, and muscovite in the schist seemingly developed by emanations from the adjacent pegmatite (Fig. 18; Pl. 2, fig. 3). Partly replaced relics of schist occur well into the pegmatite and indicate nonmobile emplacement for much of the outer portion of the body. However, near the center of the pegmatite a quartz pod containing minor muscovite occurs. The presence of this pod suggests that the central region of the pegmatite is zoned: it is an indication that the material present was mobile during emplacement in contrast to the material near the outer margins of the body. The schist foliation is contorted locally within the pegmatite, but it is uncertain whether the disturbance represents minor displacement of the host rock or pre-emplacement regional drag folding.

The pegmatite on Highway 6A has in its outer portion a foliation consisting chiefly of aligned grains of plagioclase, perthite, and quartz (Fig. 19). The alignment is parallel to the foliation of the surrounding gneiss and continuous with it. Toward the central part of the pegmatite the alignment becomes disoriented, the amount of coarse perthite increases, and near the center occur two sizable pods of coarse quartz and perthite. These pods appear to be zones and suggest mobility for the central part of this body, as at the Highway 10 pegmatite. The aligned texture in the outer part of the body was probably inherited from the gneiss. The alignment of these grains is difficult to explain by flow or settling during crystallization of the body because it is dis-

FIGURE 19.—PEGMATITE ON HIGHWAY 6A

Vertical section. Alignment of pegmatite minerals suggests nonmobility for outer part of pegmatite, but quartz-perthite pods suggest mobility for central portion. Contact between pegmatite and wall rock is generally abrupt but is gradational in some places.

cordant to the contact, or by post-emplacement deformation because it is not present through the entire pegmatite.

The wall rocks surrounding both bodies are enriched in pegmatitic constituents, and the pegmatites evidently were emplaced by introduction rather than local derivation. The constituents were perhaps introduced as a fluid or as diffusing ions that filled the site of what is now the central part of each pegmatite and then metasomatically replaced much of the surrounding wall rock. Another interpretation is that the country rocks formerly at the site of emplacement were transformed metasomatically by invading "juices" or ions, and that the pegmatized material became mobile near the present cores or centers of the bodies.

Summary and Conclusions

Any comprehensive theory of pegmatite origin must include a consideration of the processes by which the bodies were emplaced. The writer attempted a systematic analysis of the emplacement problem and classified possible mechanisms of emplacement as follows:

- Displacement: country rocks moved aside from site of emplacement
  - Forceful emplacement: country rocks moved aside by pegmatite
  - Permissive emplacement: pre-existing opening in country rock filled by pegmatite
- Nondisplacement: no movement aside of country rocks occurred
  - Mobility: material at pegmatitic site was bodily mobile during emplacement
    - Introduction: pegmatitic material introduced into site of emplacement
    - Local derivation: pegmatitic material derived locally
      - Host-rock breakdown: pegmatite derived by chemical breakdown of pre-existing rock
      - Residual segregation: pegmatite derived by crystallization, in place, of residual magmatic fluids
  - Nonmobility: material at pegmatitic site not bodily mobile during emplacement
    - Introduction
    - Local derivation

A review of the literature on emplacement reveals that certain lines of evidence are interpreted differently by different authors, whereas on other features there is more agreement. In general, criteria for emplacement need constant re-evaluation as additions are made to knowledge of the conditions of pegmatite formation.

The evidence used by the writer in considering mechanisms of emplacement and presented in the form of lists of criteria consists mainly of field features. Other reliable lines of evidence undoubtedly exist and could be applied to the problem.

A field study of a number of New England granitic pegmatites suggests that the specific mode of emplacement of a pegmatite can be determined only under very favorable circumstances. Only a few of the pegmatites studied exhibit the most truly diagnostic features, and exposures of useful features are commonly poor. However, for many of the bodies the mechanisms of emplacement could be reduced to a few major possibilities.

The various pegmatites studied by the writer have evidently been emplaced by several processes. Many individual pegmatites exhibit evidence for more than one mode of emplacement. The quantitative importance of the various mechanisms, however, is difficult to determine. Forceful emplacement was evidently significant at the Bordonaro, Chandler Mills, Ruggles, and Strickland bodies, and the pegmatite on Roaring Brook. Displacement of host rock was also important at the Kirk No. 1 and Rice-Palermo No. 26 pegmatites, but whether the displacement was forceful or permissive could not be determined from the evidence observed. Mobile introduction, in addition to either forceful or permissive emplacement, was evidently significant in the development of the Danbury, Gillette, and Wadhams-Tucker pegmatites. Mobile introduction, forceful emplacement, or permissive emplacement, or some combination of the three, probably occurred at the Britton, Case No. 1, E. E. Smith, Kilton, Marston, Nancy No. 2, and Palermo No. 1 pegmatites.

Displacement of host rock may or may not have occurred at the Coleman, Hoyt Hill, Kirk No. 2, and Paul pegmatites, and whether the emplacement was mobile or nonmobile is uncertain. The pegmatites on highways 6A and 10 did not cause significant displacement of their host rocks and developed partly by mobile and partly by nonmobile means.

References Cited

Andersen, O., 1931, Discussions of certain phases of the genesis of pegmatites: Norsk. Geol. Tidsskr., v. 12, p. 1–56

Balk, Robert, 1937, Structural behavior of igneous rocks: Geol. Soc. America Mem. 5, 177 p.

Bannerman, H. M., 1943, Structural and economic features of some New Hampshire pegmatites: Concord, N. H., N. H. State Plan. and Devel. Comm., Min. Res. Survey, pt. VII, 31 p.

Barth, T. F. W., 1952, Theoretical petrology: N. Y., John Wiley & Sons, 387 p.

Bastin, E. S., 1910, Economic geology of the feldspar deposits of the United States: U. S. Geol. Survey Bull. 420, 85 p.

—— 1911, Geology of the pegmatites and associated rocks of Maine: U. S. Geol. Survey Bull. 445, 152 p.

Bateman, A. M., 1950, Economic mineral deposits: 2d ed., N. Y., John Wiley & Sons, 916 p.

Billings, M. P., 1925, On the mechanics of dike intrusion: Jour. Geology, v. 33, p. 140–150

—— 1954, Structural geology: 2d ed., N. Y., Prentice-Hall, 514 p.

—— 1955, Geologic map of New Hampshire: Concord, N. H., N. H. State Plan. and Devel. Comm.

Birch, F., *et al.*, 1942, Handbook of physical constants: Geol. Soc. America Special Paper 36, 325 p.

Cameron, E. N., 1951, Feldspar deposits of the Bryson City district, North Carolina: N. C. Dept. Conserv. and Devel. Div. Min. Res. Bull. 62, 100 p.

Cameron, E. N., and Shainin, V. E., 1947, The beryl resources of Connecticut: Econ. Geology, v. 42, p. 353–367

Cameron, E. N., Jahns, R. H., McNair, A. H., and Page, L. R., 1949, Internal structure of granitic pegmatites: Econ. Geology, Mon. 2, 115 p.

Cameron, E. N., Larrabee, D. M., McNair, A. H., Page, J. J., Shainin, V. E., and Stewart, G. W., 1945, Structural and economic characteristics of New England mica deposits: Econ. Geology, v. 40, p. 369–393

Cameron, E. N., Larrabee, D. M., McNair, A. H., Page, J. J., Stewart, G. W., and Shainin, V. E., 1954, Pegmatite investigations, 1942–45, New England: U. S. Geol. Survey Prof. Paper 255, 352 p.

Chapman, C. A., 1941, The tectonic significance of some pegmatites in New Hampshire: Jour. Geology, v. 49, p. 370–381

—— 1952, Structure and petrology of the Sunapee quadrangle, New Hampshire: Geol. Soc. America Bull., v. 63, p. 381–426

—— 1953, Geology of the Sunapee quadrangle: Concord, N. H., N. H. State Plan. and Devel. Comm., 32 p.

Cloos, Ernst, 1937, The application of recent structural methods in the interpretation of the crystalline rocks of Maryland: Md. Geol. Survey, v. 13, p. 27–105

Crosby, W. O., and Fuller, M. L., 1897, Origin of pegmatite: Am. Geologist, v. 19, p. 147–180

Currier, L. W., 1947, Granitization and its significance as a regional metamorphic process in New England: Wash. Acad. Sci. Jour., v. 37, p. 75–86

Emmons, R. C., 1940, The contribution of differential pressures to magmatic differentiation: Am. Jour. Sci., v. 238, p. 1–21

Engel, A. E. J., and Engel, C. G., 1953, Grenville series in the Northwest Adirondack Mountains, New York: Geol. Soc. America Bull., v. 64, p. 1013–1098

Flawn, P. T., 1951, Pegmatites of the Van Horn Mountains, Texas: Econ. Geology, v. 46, p. 163–192

Fowler-Billings, K., and Page, L. R., 1942, Geology of the Cardigan and Rumney quadrangles, New Hampshire: Concord, N. H., N. H. State Plan. and Devel. Comm., 31 p.

Fowler-Lunn, K., and Kingsley, L., 1937, Geology of the Cardigan quadrangle, New Hampshire: Geol. Soc. America Bull., v. 48, p. 1363–1386

Foye, W. G., 1922, Mineral localities in the vicinity of Middletown, Connecticut: Am. Mineralogist, v. 7, p. 4–12

—— 1949, The geology of Eastern Connecticut: Conn. Geol. and Nat. Hist. Survey Bull. 74, 100 p.

Gilluly, J., *et al.*, 1948, Origin of granite: Geol. Soc. America Mem. 28, 139 p.

Goodspeed, G. E., 1940, Dilation and replacement dikes: Jour. Geology, v. 48, p. 175–195

Graton, L. C., 1906, Reconnaissance of some gold and tin deposits of the Southern Appalachians: U. S. Geol. Survey Bull. 293, p. 9–118

Greenly, E., 1923, Further researches on the succession and metamorphism in the Mona complex of Anglesey: Geol. Soc. London Quart. Jour., v. 79, p. 334–351

Griffitts, W. R., and Olson, J. C., 1953, Mica deposits of the southeastern Piedmont; Part 5, Shelby-Hickory district, North Carolina; Part 6, Outlying districts in North Carolina: U. S. Geol. Survey Prof. Paper 248-D, p. 203–293

Grout, F. F., *et al.*, 1933, Problems of the batholiths: Nat. Research Council, Div. Geol. and Geog., Ann. Rept., App. A, Exhibit B, 59 p.

Hanley, J. B., 1951, Economic geology of the Rincon pegmatites, San Diego County, California: Calif. Dept. Nat. Res., Div. Mines, Special Rept. 7 B, 24 p.

Harker, A., 1939, Metamorphism: 2d. ed., London, Methuen and Co., 362 p.

Higazy, R. A., 1949, Petrogenesis of perthite pegmatites in the Black Hills, South Dakota: Jour. Geology, v. 57, p. 555–581

Hulin, C. D., 1929, Structural control of ore deposition: Econ. Geology, v. 24, p. 15–49

Hutchinson, R. W., 1955, Regional zonation of pegmatites near Ross Lake, District of Mackenzie, Northwest Territories: Geol. Surv. Canada Bull. 34, 50 p.

Jacobson, R., and Webb, J. S., 1946, The pegmatites of central Nigeria: Geol. Survey Nigeria Bull. 17, 61 p.

Jahns, R. H., 1946, Mica deposits of the Petaca district, Rio Arriba County, New Mexico: N. Mex. Bur. Mines and Min. Res. Bull. 25, 294 p.

—— 1952, Pegmatite deposits of the White Picacho district, Maricopa and Yavapai counties, Arizona: Univ. Ariz. Bull., v. 23, no. 5, 105 p.

—— 1953, The genesis of pegmatites: Am. Mineralogist, v. 38, p. 563–598, 1078–1112

—— 1955, The study of pegmatites: Econ. Geology 50th Anniversary Volume, Pt. II, p. 1025–1130

Jahns, R. H., and Wright, L. A., 1951, Gem- and lithium-bearing pegmatites of the Pala district, San Diego County, California: Calif. Dept. Nat. Res., Div. Mines, Special Rept. 7A, 72 p.

Jahns, R. H., Griffitts, W. R., and Heinrich, E. W., 1952, Mica deposits of the Southeastern Piedmont; Part 1, General features: U. S. Geol. Survey Prof. Paper 248-A, p. 1–102

Jenks, W. F., 1935, Pegmatites at Collins Hill, Portland, Connecticut: Am. Jour. Sci., 5th ser., v. 30, p. 177–197

Kemp, J. F., 1924, The pegmatites: Econ. Geology, v. 19, p. 697–723

King, B. C., 1947, The textural features of the granites and invaded rocks of the Singo batholith of Uganda and their petrogenetic significance: Geol. Soc. London Quart. Jour., v. 103, p. 37–64

—— 1948, The form and structural features of aplite and pegmatite dikes and veins in the Osi area of the Northern Provinces of Nigeria and the criteria that indicate a nondilational mode of emplacement: Jour. Geology, v. 56, p. 459–475

Kruger, F. C., 1946a, Structure and metamorphism of the Bellows Falls quadrangle, New Hampshire and Vermont: Geol. Soc. America Bull., v. 57, p. 161–206

—— 1946b, The geology of the Bellows Falls quadrangle, New Hampshire and Vermont: Concord, N. H., N. H. State Plan. and Devel. Comm., 19 p.

Lahee, F. H., 1941, Field geology: 4th ed., N. Y., McGraw-Hill Book Co., 853 p.

Lindgren, Waldemar, 1933, Mineral deposits: 4th ed., N. Y., McGraw-Hill Book Co., 930 p.

MacGregor, M., and Wilson, G., 1939, On granitization and associated processes: Geol. Mag., v. 76, p. 193–215

Maurice, C. S., 1940, The pegmatites of the Spruce Pine district, North Carolina: Econ. Geology, v. 35, p. 49–78, 158–187

Nevin, C. M., 1949, Principles of structural geology: 4th ed., N. Y., John Wiley & Sons, 410 p.

Noble, J. A., 1952, Evaluation of criteria for the forcible intrusion of magma: Jour. Geology, v. 60, p. 34–57

Olson, J. C., 1942, Mica bearing pegmatites of New Hampshire: U. S. Geol. Survey Bull. 931-P, p. 363–403

—— 1950, Feldspar and associated pegmatite minerals in New Hampshire: Concord, N. H., N. H. State Plan. and Devel. Comm., Min. Res. Survey, Pt. XIV, 50 p.

Page, L. R., *et al.*, 1953, Pegmatite investigations, 1942–1945, Black Hills, South Dakota: U. S. Geol. Survey Prof. Paper 247, 228 p.

Perrin, R., 1956, Granite again: Am. Jour. Sci., v. 254, p. 1–18

Quirke, T. T., and Kremers, H. E., 1943, Pegmatite crystallization: Am. Mineralogist, v. 28, p. 571–580

Ramberg, H., 1949, The facies classification of rocks: a clue to the origin of quartzo-feldspathic massifs and veins: Jour. Geology, v. 57, p. 18–54

—— 1952, The origin of metamorphic and metasomatic rocks: Chicago, Univ. Chicago Press, 317 p.

—— 1955, Natural and experimental boudinage and pinch-and-swell structures: Jour. Geology, v. 63, p. 512–526

Ramberg, H., 1956, Pegmatites in West Greenland: Geol. Soc. America Bull., v. 67, p. 185–214

Rice, W. N., and Foye, W. G., 1927, Guide to the geology of Middletown, Connecticut, and vicinity: Conn. Geol. and Nat. Hist. Survey Bull. 41, 137 p.

Sederholm, J. J., 1926, On migmatites and associated pre-Cambrian rocks of Southwestern Finland; Part II, The region around the Barösundsfjärd west of Helsingfors and neighbouring areas: Comm. Geol. Finlande Bull., v. 12, No. 77, 143 p.

Shainin, V. E., 1948, Economic geology of some pegmatites in Topsham, Maine: Maine Geol. Survey Bull. 5, 32 p.

Shaub, B. M., 1940, On the origin of some pegmatites in the town of Newry, Maine: Am. Mineralogist, v. 25, p. 673–688

Spurr, J. E., and Garrey, G. H., 1908, Economic geology of the Georgetown quadrangle (together with the Empire district), Colorado, with general geology by S. H. Ball: U. S. Geol. Survey Prof. Paper 63, 422 p.

Staatz, M. H., and Trites, A. F., 1954, Geology of the Quartz Creek pegmatite district, Gunnison County, Colorado: U. S. Geol. Survey Prof. Paper 265, 111 p.

Stark, J. T., and Barnes, F. F., 1932, The structure of the Sawatch Range: Am. Jour. Sci., 5th ser., v. 24, p. 471–480

Sterrett, D. B., 1923, Mica deposits of the United States: U. S. Geol. Survey Bull. 740, 342 p.

Stockwell, C. H., 1933, Genesis of a lithium pegmatite on the Bear mineral claim, Southeast Manitoba: Roy. Soc. Canada Trans., 3rd ser., v. 27, sec. 4, p. 27–36

Stoll, W. C., 1950, Mica and beryl pegmatites in Idaho and Montana: U. S. Geol. Survey Prof. Paper 229, 64 p.

Tolman, C., 1931, Quartz dikes: Am. Mineralogist, v. 16, p. 278–299

Trueman, J. D., 1912, The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geology, v. 20, p. 228–258, 300–315

Uspensky, N. M., 1943, On the genesis of granitic pegmatites: Am. Mineralogist, v. 28, p. 437–447

Walton, M., 1955, The emplacement of granite: Am. Jour. Sci., v. 253, p. 1–18

Waters, A. C., and Krauskopf, K., 1941, Protoclastic border of the Colville batholith: Geol. Soc. America Bull., v. 52, p. 1355–1418

Webb, J. S., 1946, A replacement "pegmatite" vein in the Carn Brea granite: Geol. Mag., v. 83, p. 177–185

Wells, A. K., and Bishop, A. C., 1954, The origin of aplites: Geol. Assoc. Proc., v. 65, p. 95–114

Williams, H., Turner, F. J., and Gilbert, C. M., 1954, Petrography: San Francisco, Freeman and Co., 406 p.

The Eagle-Picher Company, Cardin, Okla.

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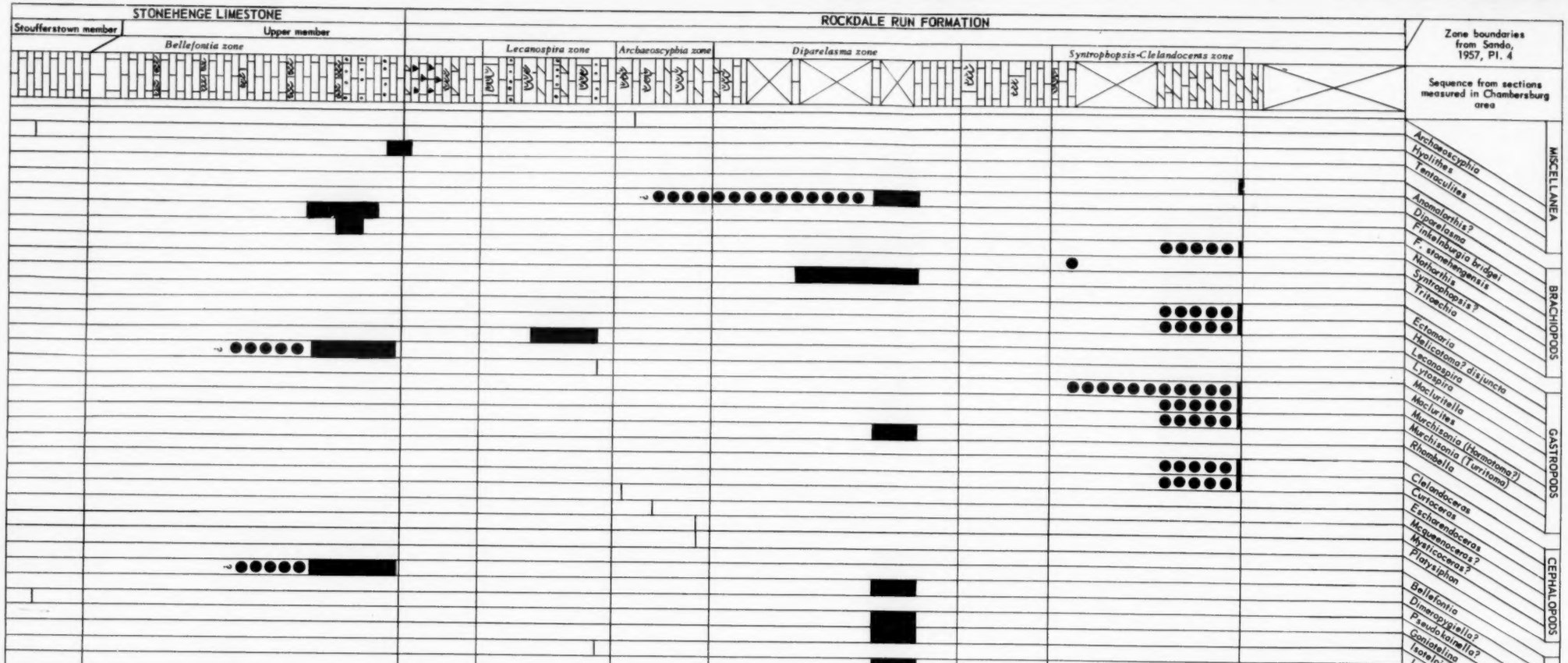
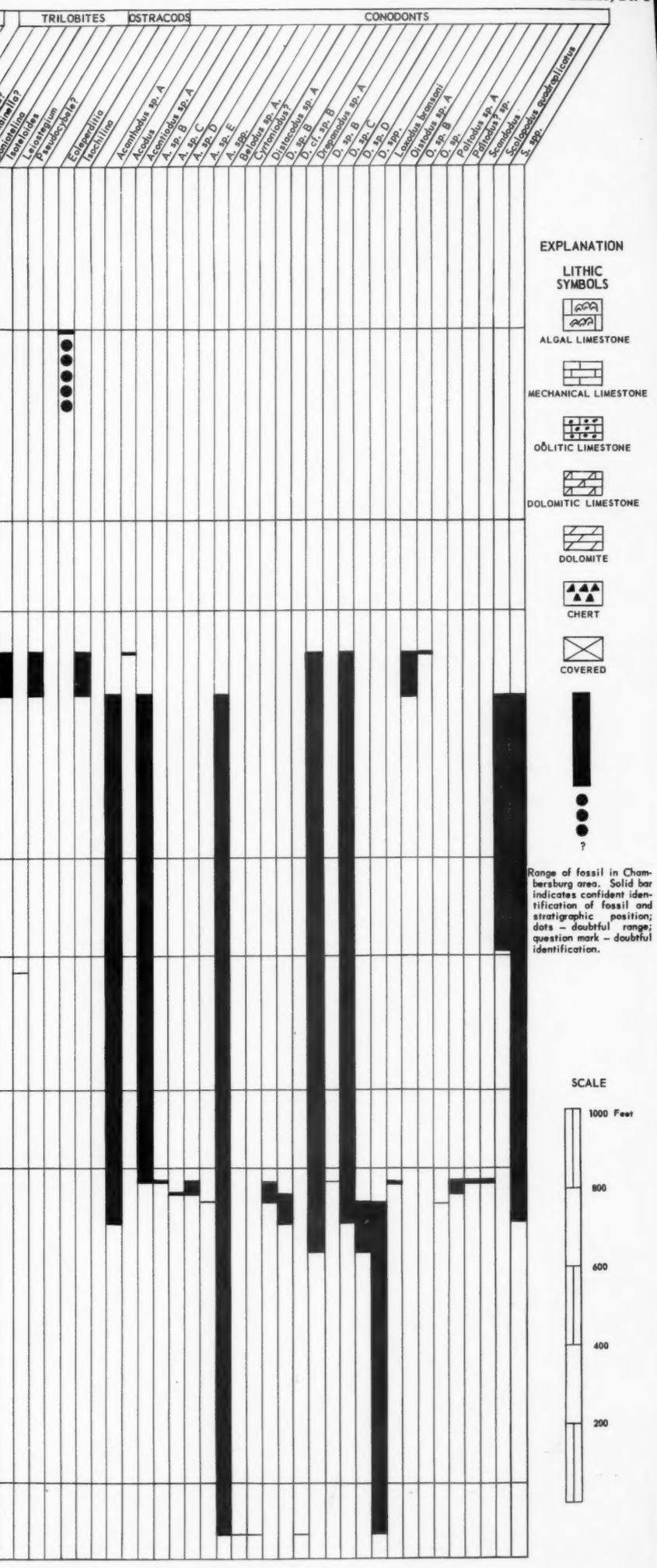
COLUMNAR SECTION AND FAUNAL ZONES





AUNAL RANGE CHART

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LOWER ORDOVICIAN SECTION NEAR CHAMBERSBURG, PENNSYLVANIA

BY WILLIAM J. SANDO

ABSTRACT

The Lower Ordovician series (Beekmantown group) is at least 3500 feet thick near Chambersburg, Pennsylvania, approximately 1200 feet thicker than was earlier estimated. Reverse faulting in the upper part of the section may have obscured the youngest Early Ordovician and oldest Middle Ordovician beds in the area, so that the total thickness of the Lower Ordovician series may be comparable to the thickness measured in Maryland.

The Beekmantown group is represented by the Stonehenge limestone and the Rockdale Run formation. Moving the base of the Stonehenge downward to include about 200 feet of strata previously referred to the underlying Conococheague formation facilitates regional correlation without impairing mappability. Newly discovered fossils from the Rockdale Run formation add notably to the faunas of the upper part of that formation. Comparison of a trilobite-brachiopod faunule near the top of the *Diparelasma* zone (Cotter-Powell equivalent) with similar fossils in Zone J of the Utah-Nevada sequence suggests that Zone J may be no younger than the Powell dolomite of the Ozark region. Difficulty in recognizing correlatives of the Smithville and Black Rock formations suggests the need for restudy of these units in the Ozark region.

Evidence from the Chambersburg section neither proves nor contradicts the hypothesis of continuous deposition from Early Ordovician to Middle Ordovician time in the central Appalachian area.

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INTRODUCTION

The section exposed near Chambersburg, Franklin County, Pennsylvania (Fig. 1), has served as a reference for Lower Ordovician stratigraphy in the Great Valley of southern Pennsylvania and Maryland since investigation of the Mercersburg-Chambersburg district by Stose (1908; 1909). E. O. Ulrich identified the fossils listed in Stose's papers, delineated faunal zones, and established time boundaries in the section. The Chambersburg section and a section near the mouth of Licking Creek in the Mercersburg quadrangle formed the basis for the stratigraphy of Ulrich's Maryland Basin (Ulrich, 1911, p. 652–655). Bassler's (1919) interpretation of the Cambrian and Ordovician stratigraphy of Maryland was shaped for the most part by concepts founded in the Chambersburg area.

In recent years, many of Ulrich's concepts of Lower Ordovician stratigraphy have been modified. Investigations in Washington County, Maryland (Sando, 1956; 1957), refined the classification and correlation of the Lower Ordovician strata. The Lower Ordo-

vician series proved to be about 1000 feet thicker than recorded by earlier geologists, and new faunal data provided a more precise basis for zonation and regional correlations. This paper presents the results of a restudy of the classic section at Chambersburg in an attempt to unify the picture of Lower Ordovician stratigraphy in the Great Valley of Maryland and southern Pennsylvania.

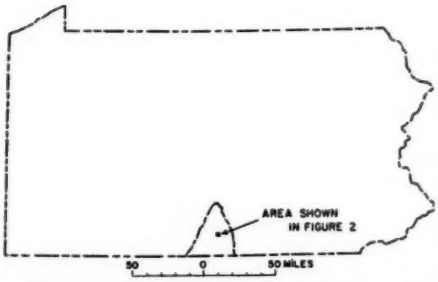

FIGURE 1.—INDEX MAP OF PENNSYLVANIA SHOWING FRANKLIN COUNTY AND AREA STUDIED

The writer spent 11 days in 1955 and 1956 mapping, measuring sections, and collecting fossils in the Chambersburg area. The original collections of Stose and Ulrich and those of the writer were studied during 1955 and 1956. Important fossils not previously described from the central Appalachian region are illustrated in Plate 2. Most of the other fossils listed are described and illustrated in papers by Sando (1957), Ulrich and Cooper (1938), Flower (1956), and Ulrich *et al.* (1942). Plate 1 shows the stratigraphic distribution of important faunal elements. Location of geologic sections and fossil localities is shown in Figure 2.

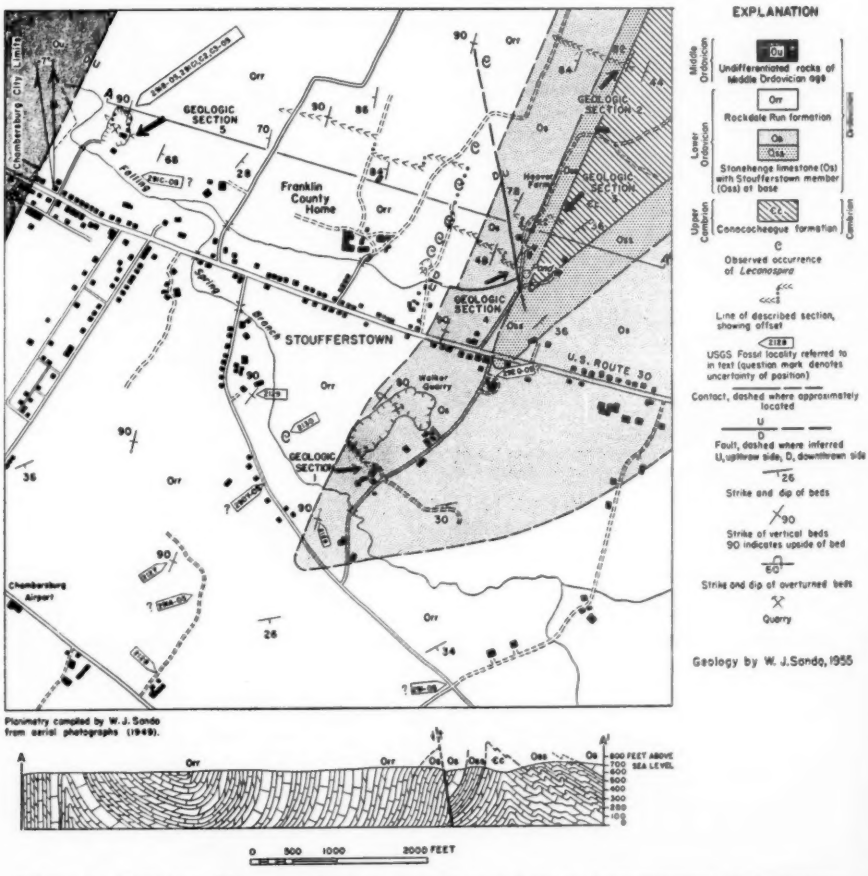
EXPLANATION
Undifferentiated rocks of Middle Ordovician age
Rockdale Run formation
Stonehenge limestone (Os) with Stoufferstown member (Oss) at base
Conococheague formation
Observed occurrence of Lecanospira
Line of described section, showing offset
USGS Fossil locality referred to in text (question mark denotes uncertainty of position)
Contact, dashed where approximately located
Fault, dashed where inferred U, upthrow side, D, downthrown side
Strike and dip of beds
Strike of vertical beds 90 indicates upside of bed
Strike and dip of overturned beds
Quarry
Geology by W. J. Sando, 1955
Chambersburg City Limits
GEOLOGIC SECTION 5
GEOLOGIC SECTION 2
GEOLOGIC SECTION 3
GEOLOGIC SECTION 4
GEOLOGIC SECTION 1
Franklin County Home
Falling Spring Branch
STOUFFERSTOWN
U.S. ROUTE 30
Walker Quarry
Chambersburg Airport
Planimetry compiled by W. J. Sando from aerial photographs (1949).
A
A'
800 FEET ABOVE SEA LEVEL
0 500 1000 2000 FEET

FIGURE 2.—GEOLOGIC MAP AND SECTION OF THE AREA EAST OF CHAMBERSBURG, PENNSYLVANIA

Acknowledgments

The facilities of the U. S. Geological Survey were used in preparation and study of the fossils. The original collections of Stose and Ulrich were made available to the writer by the U. S. National Museum. The writer has benefited greatly from the advice of R. J. Ross, Jr., G. A. Cooper, J. B. Knight, E. L. Yochelson, Franco Rasetti, and A. R. Palmer concerning identification of some of the fossils. He is particularly indebted to W. H. Hass and J. M. Berdan for identification of the conodonts and ostracods, respectively. Technical criticism by R. B. Neuman, Ross, Cooper, and Hass aided in clarification of ideas expressed in the report. N. A. Shupe photographed the fossils illustrated in Plate 2.

Stratigraphy

Stonehenge Limestone (Emended)

DEFINITION: Stose (1908, p. 703–704) named the Stonehenge limestone after the village of Stonehenge at the eastern edge of Chambersburg. This village does not appear on recent topographic or road maps; the area is now included in Stoufferstown. The Stonehenge was originally defined as a member of the Beekmantown limestone, which included all the Lower Ordovician strata. It is here given formation status in conformity with the writer's earlier classification in Washington County, Maryland (Sando, 1956; 1957).

The base of the Stonehenge limestone was originally placed at the contact of pure, fine-grained, massive limestones with underlying sandy, conglomeratic, thin-bedded limestones included in the upper part of the Conococheague formation. The Cambrian-Ordovician boundary was placed at this contact. The lower thin-bedded limestones form a persistent stratigraphic unit in southern Pennsylvania, Maryland, and West Virginia and bear a small but diagnostic Lower Ordovician trilobite assemblage. These beds are here named the Stoufferstown member and are classified as the base of the Stonehenge rather than the top of Conococheague formation. This arrangement is a better expression of temporal unity and is no less useful in geologic mapping. Beds between the top of the Stoufferstown member and the top of the Stonehenge will be referred to as the upper member. In Maryland, beds previously called "upper member of the Conococheague formation" by the writer (Sando, 1957, p. 15, 16) are now referred to the Stoufferstown member. The upper member of the Stonehenge as defined in this paper is equivalent to the upper and lower members of Maryland (Sando, 1957, p. 19).

The top of the Stonehenge was not specifically defined by Stose. He apparently based this boundary on a change from thin-bedded, coarse-grained, noncherty limestone in the upper Stonehenge to massive, fine-grained, cherty limestone in the lower part of the unnamed remainder of his Beekmantown formation. The definition used by the writer is based on an increase in dolomite in the upper part of the Beekmantown group. Studies in Maryland indicate fairly persistent dolomite beds at approximately the same horizon where the earlier geologists drew the top of the Stonehenge. Chert in residual soils developed from the lower part of the overlying Rockdale Run formation (cryptozöon chert zone of Sando, 1957, p. 21) aids in placement of the contact in areas of poor bedrock exposure. In the Chambersburg area, a bed of dolomitic limestone identified in Geologic Sections 1 and 4 marks the top of the Stonehenge; in Geologic Section 2, the top is covered but can be placed approximately under the cryptozöon chert zone.

Stose did not designate a type section for the Stonehenge limestone, and none of the sections measured by the writer gives a complete picture of lithic and faunal succession in the formation. The present concept of the Stonehenge is based on a composite section made up of the four measured. Geologic Section 3 is the type section of the Stoufferstown member.

LITHIC CHARACTER: Two genetic types of limestone characterize the Stonehenge: algal limestone and mechanical limestone (*see* Sando, 1957, p. 33–39 for detailed discussion of these terms). Massive, fine-grained algal beds compose nearly half the formation in Maryland but are rare in the Chambersburg area. Massive beds questionably identified as algal near the base of the upper member lack the gymnosolenid bioherms and calcarenite-filled channels that characterize these beds in the Maryland sections (Sando, 1957, p. 19). The mechanical limestone is predominant in the Stonehenge of the Chambersburg area. This is characterized by thin-bedded carbonate detritus ranging in size from clay to pebbles with minor terrigenous matter in the form of thin, silty, yellowish-weathering seams, some of which weather in relief. Two lithofacies may be distinguished within the mechanical limestones: (1) The fine-grained lithofacies is com-

posed dominantly of carbonate detritus with component particles ranging from clay to sand in size. Oöliths are uncommon in this facies. The yellowish-weathering seams in these rocks are predominantly fine-grained dolomite with subordinate amounts of silt-sized quartz grains; these seams do not commonly stand in sharp relief on a weathered surface. (2) The coarse-grained lithofacies is composed dominantly of carbonate detritus with particles ranging from clay to cobble size; tabular limestone fragments commonly 2–3 inches in diameter characterize this facies. Many of these beds are oölitic. The yellowish-weathering seams have a higher terrigenous quartz content and in places are reinforced by secondary silica, so that in many places they weather in sharp relief. Bright-orange-weathering patches of dolomite half an inch or less in diameter are a distinctive feature.

The most persistent development of the coarse-grained lithofacies is the Stoufferstown member, which is almost entirely this rock type. The member expresses itself as a prominent ridge because of the resistant nature of this lithofacies. Approximately the lower half of the upper member of the Stonehenge is fine-grained mechanical limestone with a few interbeds of algal limestone. In the upper half of this member, the coarse-grained and fine-grained lithofacies of mechanical limestone are interbedded in approximately equal quantities. A few beds or lenses of algal limestone also occur in this part of the member.

THICKNESS: The total thickness of the Stonehenge is approximately 1000 feet. The Stoufferstown member is 217 feet thick in its type section (Geologic Section 3). The upper member cannot be measured accurately within the area studied because of faulting and poor exposures of the top, but about 692 feet was measured in Geologic Section 2, where the top is covered. A composite section made by comparing the four incomplete sections measured suggests that the upper member is about 800 feet thick.

FAUNA AND CORRELATION: Fossils are rare in the Stoufferstown member, but a thin bed of oölitic limestone 62 feet above the base yielded:

Hyolithes sp. (Pl. 2, fig. 15)
Finkelnburgia sp.
Inarticulate brachiopod fragments
Ophiletid gastropod
Orthoceracone cephalopod
Pseudokainella? aff. *P.? armatus* (Pl. 2, figs. 31, 32)
Merostome(?) fragments
Acontiodus sp.
Belodus sp. *A*. (Pl. 2, fig. 23)
Cyrtoniodus? sp. (Pl. 2, fig. 26)
Distacodus cf. sp. *B*
Drepanodus sp.

Trilobite fragments and an indeterminate conodont were also found about 100 feet below the top. In Maryland, *Clelandia* sp. and *Symphysurina* cf. *S. woosteri* were collected about 30 feet below the top of the member (Sando, 1957, p. 16).

Symphysurina and *Clelandia* characterize a zone near the base of the Lower Ordovician series; one or both of these genera have been reported from the Tribes Hill limestone of New York (Raymond, 1937, Pl. 1, fig. 25), the Oneota dolomite of Minnesota and Wisconsin (Palmer *in* Ross, 1951, p. 116–117; Ulrich *in* Walcott, 1925, p. 115–116), Zones A, B, and C of the Garden City formation of northeastern Utah (Ross, 1951, p. 115–117), the *Symphysurina* zone (Zone B of Ross) of the Pogonip group of western Utah and eastern Nevada (Hintze, 1952, p. 6–7), the Goodwin limestone of Nevada (Cooper *in* Nolan *et al.*, 1956, p. 26–27), the Mons formation of Alberta and British Columbia (Walcott, 1925, p. 112–113), the Deadwood and Pilgrim formations of Montana and Wyoming (Lochman and Duncan, 1950, p. 351, 352), and the Manitou formation of Colorado (Bass and Northrop, 1953, p. 908–911). Thus, the age and correlation of the uppermost part of the Stoufferstown member are well established.

The presence of *Pseudokainella?* and conodonts indicates that the lower part of the member is also Lower Ordovician. The trilobite questionably referred to *Pseudokainella* is similar to *P.? armatus* Hintze (1952, p. 218, Pl. 5, figs. 1–5), which occurs in Hintze's *Symphysurina* zone (Zone B of Ross, 1951) of the Pogonip group of western Utah. This species also occurs in the lowest beds of the Garden City formation in northern Utah (Ross, Personal communication, 1957). According to W. H. Hass (Personal communication, 1957), the conodonts of the lower Stoufferstown suggest an Early Ordovician age. Of the five genera in this zone, only *Belodus* and *Cyrtoniodus?* do not occur in higher beds in the Chambersburg section.

The exact position of the Cambrian-Ordovician boundary is uncertain. In Maryland, Trempealeau trilobites (*Plethometopus*) occur up to about 400–500 feet below *Symphysurina* and *Clelandia* (Sando, 1957, p. 16). Wilson

(1952, p. 305, 307, 316) collected *Prosaukia*, *Plethometopus*, *Stenopilus*, *Leiocoryphe*, and *Bowmania* from beds near the top of the Conococheague formation about 1.5 miles northeast of the type locality of the Stonehenge limestone. A collection made by the writer from USGS Locality 2186 about half a mile north of Wilson's locality includes *Plethometopus* and *Stigmametopus?*. Interpretation of the geologic structure suggests that this collection occurs no more than 300 feet below the base of the Stoufferstown member. Saukid trilobites were collected from USGS Locality 2187 (1100 feet southeast of 2186) no more than 300 feet below the *Plethometopus* collection. Thus, the Cambrian-Ordovician boundary may lie somewhere within the 300-foot unfossiliferous interval between the *Plethometopus* bed of the Conococheague formation and the *Pseudokainella?* bed of the Stoufferstown member. For convenience, the Cambrian-Ordovician boundary is tentatively placed at the base of the Stoufferstown member.

In the type area, the upper member of the Stonehenge limestone has yielded few fossils. As in the Maryland sections, *Bellefontia collieana*, *Finkelnburgia bridgei*, *F. stonehengensis*, and lytospirid gastropods are most abundant. *Tentaculites lowdoni* (Pl. 2, fig. 16) occurs near the top of the Stonehenge and ranges into the lowest 20 feet of the overlying Rockdale Run formation. Conodonts are common in oölitic limestones in the upper 200 feet of the member. The conodont assemblage includes species of *Acontiodus*, *Acanthodus*, *Distacodus*, *Drepanodus*, *Loxodus*, *Oistodus*, *Paltodus*, *Scandodus*, and *Scolopodus*. Five of the nine genera range into the Rockdale Run formation, but many of the species seem to be restricted to the Stonehenge (Pl. 1). One species, *Loxodus bransoni* (Pl. 2, fig. 17), was previously described from the Oneota dolomite in Iowa and Minnesota (Furnish, 1938, p. 339).

A more extensive fauna (excluding the conodonts) was collected from the same interval in Maryland (Sando, 1957, p. 20). The Stonehenge limestone as here emended is correlated with the Gasconade dolomite of the Ozark region and its equivalent formations in North America. The writer (Sando, 1957, p. 20–21, Pl. 3) recently discussed these correlations

Rockdale Run Formation

DEFINITION: The Rockdale Run formation (Sando, 1956; 1957, p. 21–28) was named from exposures along Rockdale Run near Hicksville, Washington County, Maryland, where it constitutes the interval between the top of the Stonehenge limestone and the base of the Pinesburg Station dolomite. Although the top of the formation is not exposed in the Chambersburg section, lithic and faunal similarity with the Maryland sections justifies extension of the name to the Chambersburg region.

The upper 880 feet of the Rockdale Run formation is almost completely covered in Geologic Section 4. Structural interpretations (Fig. 2) suggest that the base of Geologic Section 5 correlates with a horizon about 250 feet above the base of Unit 53 of Geologic Section 4. The following remarks pertain to the composite section constructed in this manner.

LITHIC CHARACTER: The Rockdale Run formation consists of a thick series of interbedded limestones and dolomites. The distribution of lithic types in the Chambersburg section is approximately the same as that of the Maryland sections of the formation (Sando, 1957, p. 21, 22): algal and mechanical limestones dominate the lower two-thirds of the formation; dolomites and dolomitic limestones are abundant in the upper third. The percentage of dolomite does not seem as high as that of the Maryland sections, but exposures are too incomplete for exact comparison. Most of the magnesium seems to be in dolomitic limestones rather than in dolomites, and the percentage of purer limestones seems to be higher, even in the upper third of the formation. Thus, the eastward increase in the limestone/dolomite ratio (Sando, 1957, p. 22–24) may be more pronounced in southern Pennsylvania than in Maryland.

THICKNESS: The formation measures 2547 feet in Geologic Section 4, but the upper 880 feet is almost completely covered. The top of the formation is not exposed. The measured thickness compares favorably with complete measurements of the formation in Maryland (Sando, 1957, p. 24).

FAUNAS AND CORRELATION: The distribution of fossils in the Rockdale Run formation in the Chambersburg section (Pl. 1) fits well into the writer's scheme of faunal zones for Maryland sections of the formation (Sando, 1957, p. 24–25, Pl. 4). Inasmuch as the faunas and correlation of these zones have been discussed previously (Sando, 1957, p. 24–28, Pl. 3), the following discussion is confined to new faunal data and their significance in correlation with other formations in North America.

Although correlation of the *Lecanospira* zone is not altered, two new faunal elements, *Leiostegium* cf. *L. puteatum* (Pl. 2, figs. 28, 29) and *Acontiodus* sp. *A* (Pl. 2, fig. 25), are added. *Leiostegium* was known previously for the most part from beds thought to be equivalent to some part of the Stonehenge limestone; it was reported from this interval in Vermont, Quebec, Nevada, Utah, and British Columbia (Hintze, 1952, p. 188–189). *Leiostegium puteatum* Raymond (1924, p. 455, Pl. 14, figs. 12, 13, 16, 19) is known only from the Highgate slate of Vermont, a formation of Stonehenge age (Twenhofel *et al.*, 1954, Pl. 1). According to R. J. Ross, Jr. (Personal communication, 1957), *L. manitouensis*, which occurs in the lower part of the Manitou formation of Colorado and in Zone D of the Garden City formation and Pogonip group in Utah and Nevada, cannot be distinguished from *L. puteatum*. The genus ranges into Zone E, which Ross (1951, p. 32) questionably correlated with the Roubidoux formation. Ross (Personal communication, 1957) also reports that

> "*Leiostegium* occurs in the uppermost beds of the Goodwin formation in the Antelope Range of Nevada; the upper Goodwin may range as high as Zone G."

Hence, the occurrence of the genus in the *Lecanospira* zone of Pennsylvania may not constitute an extension of its stratigraphic range. The conodont *Acontiodus* sp. *A* has a long range in the Chambersburg section but is not known outside the Chambersburg area.

The *Archaeoscyphia* zone is represented in the Chambersburg section by fossils typical of this zone in Maryland. An addition to the fauna is *Scolopodus quadraplicatus* (Pl. 2, fig. 21), which occurs in the Jefferson City dolomite of Missouri (Branson and Mehl, 1933, p. 63) and the Shakopee dolomite of Wisconsin and Minnesota (Furnish, 1938, p. 332–334). The occurrence of this conodont in the Shakopee suggests altering the previous correlation of the *Archaeoscyphia* zone (Sando, 1957, Pl. 3) to include a Shakopee equivalent.

An assemblage of fossils here designated the *Isoteloides* faunule from its most abundant species, *Isoteloides* cf. *I flexus*, has been discovered near the top of the *Diparelasma* zone at three localities in the Chambersburg area and in a collection (USGS Locality 292X-OS) made by Ulrich and Stose near the mouth of Licking Creek, Pennsylvania (Stose, 1909, p. 7). Important fossils collected from these beds are:

Diparelasma cf. *D. marylandicum*
Tritoechia sp.
Rhombella sp. (Pl. 2, figs. 3, 4)
Isoteloides cf. *I. flexus* (Pl. 2, figs. 33–35)
Goniotelina sp. (Pl. 2, fig. 37)
Dimeropygiella? sp. (Pl. 2, fig. 30)
Pseudocybele? sp. (Pl. 2, fig. 36)
Isochilina sp. (Pl. 2, fig. 10)
Acodus sp. (Pl. 2, fig. 24)
Acontiodus sp. *A* (Pl. 2, fig. 25)
Acanthodus sp. *A* (Pl. 2, fig. 20)
Drepanodus sp. *A* (Pl. 2, fig. 27)
Drepanodus sp. *C*
Oistodus sp. *A* (Pl. 2, fig. 19)
Oistodus sp. *B*
Scolopodus quadraplicatus (Pl. 2, fig. 21)

Rhombella was previously known only from formations correlative with the *Lecanospira* zone (Bridge and Cloud, 1947, p. 555); its occurrence in the Chambersburg section extends its stratigraphic range. *Isochilina*, although common in the upper part of the Lower Ordovician series, has a long range extending into the Silurian; until more is known about Lower Ordovician species, it is of little value for correlation. The association of the trilobites *Isoteloides*, *Goniotelina*, *Dimeropygiella?*, and *Pseudocybele?* and the brachiopods *Diparelasma* and *Tritoechia* suggests correlation of the *Isoteloides* beds with Zone J of the Garden City formation and Pogonip group of Utah and Nevada (Ross, 1951; Hintze, 1952). Possible representatives of the Cassin fauna of Vermont (Cady, 1945, p. 545–546) are *Isoteloides*, *Goniotelina*, *Diparelasma*, and *Isochilina*. The presence of *Scolopodus quadraplicatus* supports the previous correlation of part of the *Diparelasma* zone (Sando, 1957, p. 27, Pl. 3) with the Cotter and Powell dolomites of Missouri and the Shakopee dolomite of the Upper Mississippi Valley; the other conodont species have not been recorded outside the Chambersburg area.

The Chambersburg section yielded the following fossils not found in Maryland exposures of the *Syntrophopsis-Clelandoceras* zone:

Anomalorthis? cf. *A. vermontensis* (Pl. 2, figs. 11–14)
Nothorthis pennsylvanica Ulrich and Cooper (1938, p. 107, Pl. 17a, figs. 1–3)
Ectomaria sp. (Pl. 2, fig. 2)
Helicotoma? disjuncta (Pl. 2, fig. 8)
Maclurites small spp.
Murchisonia (Turritoma) cf. *acrea* (Pl. 2, fig. 1)
Curtoceras internastriatum Ulrich, Foerste, Miller, and Furnish (1942, p. 72–73, Pl. 39, fig. 2)
Eoleperditia sp. (Pl. 2, fig. 9)

Anomalorthis characterizes the Whiterock stage of Cooper (1956, p. 7–8, Chart 1), the lowest zone in the Middle Ordovician of North America, but *Anomalorthis vermontensis* Ulrich and Cooper (1938, p. 129–130; Pl. 22*B*, figs. 7–10) differs from all other species referred to the genus in lacking a thickened muscular platform in the pedicle valve; thus, the generic reference of this species is in doubt. Brachiopods from the Chambersburg section differ from *Anomalorthis? vermontensis* only in shell proportions. *A. vermontensis* is known only from the Providence Island formation of Ulrich (*in* Ulrich and Cooper, 1938, p. 26), which corresponds to Division E of Brainerd and Seely (1890) in the Shoreham, Vermont, section. These strata were called Bridport dolomite by Cady (1945, p. 545–546). It would seem that *Anomalorthis? vermontensis* is a progenitor of the Middle Ordovician *Anomalorthis* stock. *Nothorthis* has been reported from Zone K of the Garden City formation in Utah (Ross, 1951, p. 27) and the Lower Ordovician section at Lèvis, Quebec (Ulrich and Cooper, 1938, p. 107); a single valve from the Cassin limestone in Vermont may belong to this genus (Ulrich and Cooper, 1938, p. 107). *Nothorthis* is also known from Middle Ordovician beds in Alabama (Cooper, 1956, p. 315–316). *Curtoceras internastriatum* occurs in the Cassin (Ulrich *et al.*, 1942, p. 72–73). *Helicotoma? disjuncta* is known only from Canada, but a similar species, *H.? similis*, occurs in the Cassin (Division E) (Whitfield, 1890, p. 31). The remaining faunal elements are characteristic of rocks of late Early Ordovician age, but their distribution is too poorly known for exact correlation.

PROBLEMS OF CORRELATION OF THE UPPER FAUNAS: The *Syntrophopsis-Clelandoceras* zone was previously correlated with the Smithville formation of Arkansas (Sando, 1957, p. 27, Pl. 3). Fossils that formed the basis for this correlation are *Syntrophopsis* cf. *S. minor* and *Clelandoceras* sp. *Syntrophopsis minor* is known only from the Cotter and Smithville formations (Ulrich and Cooper, 1938, p. 236). *Clelandoceras* was recently reported from the Chepultepec dolomite of Virginia (Unklesbay and Young, 1956, p. 487) but is not known from formations of late Early Ordovician age other than the Smithville. The *Ceratopea* species found in the *Syntrophopsis-Clelandoceras* zone in Maryland (Sando, 1957, p. 127–128, Pl. 11, figs. 24–26) proved to be *Ceratopea unguis*, a Smithville species (Yochelson and Bridge, 1957, p. 300–301, Fig. 103). Evidence from the Chambersburg section does not appear to contradict the previous correlation of the *Syntrophopsis-Clelandoceras* zone. This zone and the upper part of the *Diparelasma* zone seem to be correlative with the Cassin fauna of Vermont, which contains equivalents of the Cotter, Powell, and Smithville formations of the Ozark region (Cady, 1945, p. 545–546, Fig. 3).

The newly discovered trilobite-brachiopod assemblage (*Isoteloides* faunule) near the top of the *Diparelasma* zone apparently correlative with Zone J of the Garden City formation and the Pogonip group presents a problem. Fossils from the *Diparelasma* zone in Maryland suggest that the zone is no younger than the Powell dolomite of the Ozark region; evidence from the Chambersburg section supports this correlation. However, Ross (1951, p. 31–32) implied that his Zone J may be of Black Rock age or younger, a reasonable suggestion on the basis of evidence available to him. Ross based his correlation mainly on the Middle Ordovician affinities of the trilobite fauna of Zone J. *Kawina* occurs in the Middle Ordovician Chazy limestone and younger beds, and *Carolinites* is known from lower Middle Ordovician beds. Later work by Hintze (1952, p. 145) revealed that *Carolinites* occurs as low as Zone G2 in the western Utah section; Zone G2 was previously thought by Ross (1951, p. 32) to correlate with the Rich Fountain formation of Cullison (1944) in the Ozark region. Furthermore, Hintze (1952, p. 18) recorded *Leseuerilla?* sp., *Campbelloceras?* sp., and *Catoraphiceras* sp. from Zone J. *Leseuerilla* ranges from Theodosia to Powell in the Ozark region (Cullison, 1944, Pl. 2). *Campbelloceras* occurs in beds of Cotter and Powell age (Ulrich *et al.*, 1942, p. 43–47). *Catoraphiceras* is best known from the Cotter dolomite of Missouri and its equivalents (Ulrich *et al.*, 1944, p. 77), although it occurs as high as the Odenville limestone of Alabama. The Zone J trilobites referred to *Goniotelus* by Hintze (1952, p. 158–161) were placed in *Goniotelina* by Whittington (1953, p. 666); *Goniotelina* occurs in the Cotter and Powell formations (Whittington, 1953, p. 666) and may be restricted to the Cotter-Powell interval in North America (Twenhofel *et al.*, 1954, Pl. 1). Most of the other trilobite genera in Zone J have not been reported outside the Utah-Nevada region. The Zone J brachiopods *Diparelasma*, *Hesperonomia*, *Syntrophopsis*, and *Tritoechia* are exclusively Lower Ordovician forms whose ranges are compatible with a Cotter-Powell correlation (Twenhofel *et al.*, 1954, Pl. 1). Thus, the new evidence appears

to favor correlation of Ross' Zone J with part of the Cotter-Powell interval in the standard Ozark section and the upper part of the *Diparelasma* zone of the central Appalachian section.

Correlation of the Appalachian and Ozark sections with the western sections is difficult because of dissimilarity of the known faunas of these regions. Few trilobites have been found in the upper part of the eastern Lower Ordovician sections, but in Utah and Nevada, trilobites constitute most of the described faunas. Mollusks, on the other hand, are apparently more abundant in the eastern sequence. Brachiopods seem to constitute approximately the same proportion of the faunas in both eastern and western sequences. Possibly some of these faunal differences can be explained by differences in preservation of the various groups in the two sequences. In the western sequence, trilobites are commonly silicified and thus relatively easy to collect, whereas in the eastern sequence, silicified trilobites are rare. Perhaps not enough emphasis has been placed on collecting less easily recoverable fossils in both regions. The eastern sections have been studied for many years, however, and one would expect that a representative sample of the true faunal content has been collected. If the faunal differences are real, and if the revised correlations suggested above are correct, geologists working on the western sections have placed too much emphasis on the trilobites as regional stratigraphic indices. Possibly the Chazy trilobite genera were living in the western region during Cotter-Powell time and did not reach the eastern United States until Middle Ordovician time.

If Zone J is correlative with the upper part of the *Diparelasma* zone (Cotter-Powell equivalent), then one might expect to find equivalents of the *Syntrophopsis-Clelandoceras* zone (Smithville equivalent) in Zone K and/or Zone L. The Zone K fauna consists of *Nothorthis* sp., *Diparelasma* sp., *Hesperonomia* sp., and *Blastoidocrinus* cf. *B. carchariaedens* (Ross, 1951, p. 27). The Zone L fauna consists of *Anomalorthis* n. sp., *Rhynchocamara* n. sp., *Syntrophopsis transversa*, *Orthis* aff. *O. subalata*, and *Blastoidocrinus* cf. *B. carchariaedens* (Ross, 1951, p. 27). Tentative correlation of Zone K with the *Syntrophopsis-Clelandoceras* zone is suggested by the presence of *Nothorthis* in both zones. The other Zone K brachiopods, *Diparelasma* and *Hesperonomia*, have not been found in the *Syntrophopsis-Clelandoceras* zone, but their stratigraphic ranges are compatible with this correlation (*See* Twenhofel *et al.*, 1954, Pl. 1). Comparison of Zone L with the *Syntrophopsis-Clelandoceras* zone is just as tenuous. *Syntrophopsis* occurs in both zones but is known from older beds (Twenhofel *et al.*, 1954, Pl. 1). The presence of the *Anomalorthis* lineage in both zones may indicate partial contemporaneity. *Rhynchocamara* is not known in the *Syntrophopsis-Clelandoceras* zone but occurs in the Smithville formation. *Orthambonites* (represented by "*Orthis* aff. *O. sublata*" in Ross' list) ranges from upper Lower Ordovician (upper Pogonip) well into Middle Ordovician (Cooper, 1956, Chart 1, p. 294–314). *Blastoidocrinus*, which occurs in Zones K and L, is the only Chazy element whose known stratigraphic range would have to be extended downward.

Although there is a suggestion of Smithville equivalents in Zones K and L, complete equivalency of these beds with the Smithville formation does not seem to be supported by recent analysis of the brachiopods of the overlying Zone M (Kanosh shale of Hintze, 1952). According to Cooper (1956, p. 130), the Kanosh shale appears to be equivalent to the Oil Creek formation in Oklahoma and Cooper's *Anomalorthis* zone in Nevada, beds which Cooper (1956, Chart 1) places near the top of his Whiterock stage (early Middle Ordovician). Ross (1951, p. 33) also noted great similarity between the trilobite faunas of Zone M and the Oil Creek and Joins formations of Oklahoma. The writer sees no reason to challenge the Middle Ordovician age determination of Zone M. Thus, if all of Early Ordovician time is represented in the Utah-Nevada sequence and if Zone M is not earliest Middle Ordovician, then latest Early Ordovician time (Black Rock) and earliest Middle Ordovician time (lower Whiterock of Cooper) should also be represented in Zones K and/or L. The evidence at hand does not permit identification of a Black Rock equivalent in the western sequence. Possibly the source of this difficulty is in classical interpretation of the upper part of the standard Ozark section. Very little information has been published concerning faunal succession in and stratigraphic relationships of the Smithville and Black Rock formations. Possibly these formations are partial time equivalents. Until these formations and their faunas are studied in detail and the classical concepts reappraised, correlation of the upper part of the North American Lower Ordovician sequences cannot be confidently made.

Position of Lower Ordovician-Middle Ordovician Boundary

In Maryland, the Rockdale Run formation is succeeded by the Pinesburg Station dolomite, which has not yielded fossils (Sando, 1957, p. 31). The Pinesburg Station dolomite is overlain by the Row Park limestone of Neuman (1951, p. 278–286), which correlates with the middle and upper Chazy limestone of New York. Thus, the inter-epoch boundary lies somewhere in the 700–800-foot interval between the top of the *Syntrophopsis-Clelandoceras* zone of the Rockdale Run formation and the base of the Row Park limestone. The lack of unambiguous physical evidence of an appreciable hiatus in this interval in Maryland suggests continuous deposition from Early Ordovician to Middle Ordovician time (Sando, 1957, p. 28–32). The writer had hoped that the Chambersburg section might throw some light on this problem, inasmuch as Stose (1909, p. 6, geologic map) indicated lower Middle Ordovician beds (Stones River limestone) in this section.

The area immediately east of Chambersburg mapped as Stones River limestone by Stose is almost completely covered, and the writer was unable to identify beds of early Middle Ordovician age in place within it. However, a collection of fossils (USGS Loc. 290Z-0S) made by H. E. Dickhaut from a stone fence west of the quarry in which Geologic Section 5 was measured contains a silicified gastropod operculum tentatively identified as *Maclurites magnus* and a shell that may belong to *Palliseria*. *Palliseria* and large *Maclurites* opercula are characteristic of a zone in Cooper's Whiterock stage in Nevada (Cooper, 1956, p. 126–127). Unfortunately, the shell referred to *Palliseria* is too poorly preserved for certain identification, and the writer was unable to determine the geographic and stratigraphic position of the collection. The only other evidence of Middle Ordovician rocks found by the writer is a few loose blocks of Chambersburg limestone in a rock pile southeast of Falling Spring Branch about 400 feet southeast of the quarry in which Geologic Section 5 was measured. These blocks seem to have been removed from excavations in a housing development several hundred feet west of the rock pile. The apparent absence of the Pinesburg Station dolomite, probable position of the Chambersburg limestone, and scarcity of outcrops suggest that reverse faulting may have obscured stratigraphic relationships in this part of the area. The solution of this problem requires detailed mapping in an area much larger than that investigated by the writer. In any event, the Chambersburg section provides no further information on the position and nature of the Lower Ordovician-Middle Ordovician boundary in southern Pennsylvania.

Geologic Sections

Each section is located on Figure 2. Except where noted, thicknesses of units were measured with a steel tape stretched horizontally along the traverse; thicknesses were computed where bedding is not vertical. The lithic classification is that of the writer (Sando, 1957, p. 33–45) and is discussed in the text. Fossils represented in collections of the U. S. Geological Survey are designated by asterisks; other faunal identifications were made in the field. USGS locality numbers are listed at the end of the description of each unit.

Geologic Section 1—Stonehenge Limestone (Upper Member In Part)

Section begins at fence corner at southeast corner of Walker Quarry just south of Stoufferstown and extends across south end of quarry. About upper half of upper member measured; lower beds poorly exposed at northeast end of quarry.

Unit	*Description*	*Thickness (feet)*
1	Limestone, mechanical, coarse-grained lithofacies, a few thin oölitic zones; *Finkelnburgia stonehengensis** 62 and 85 feet above base; *Acanthodus* sp. *A**, *Distacodus* sp. *B**, *Drepanodus* sp. *C**, *Drepanodus* sp. *D**, *Scolopodus* sp.*, and inarticulate brachiopod fragments* 85 feet above base; small indeterminate gastropods* 33 feet above base; asaphid trilobites throughout (USGS Locs. 2102, 2104, 2105).	95.0
2	Covered; deep cleft in quarry wall occupied by water pump.	12.0
3	Limestone, mechanical, fine-grained lithofacies, zone of algal bioherms 3 feet above base; *Finkelnburgia stonehengensis** and an indeterminate brevicone cephalopod* 3 feet above base, asaphid trilobites throughout (USGS Loc. 2106). (Traverse offset 73 feet north along strike.)	26.0
4	Limestone, algal, channeled; *Bellefontia* sp.	2.0
5	Limestone, mechanical, fine-grained lithofacies; *Lytospira* sp.	13.0
6	Limestone, mechanical, oölitic; *Lyto-*	2.0

Unit	Description	Thickness (feet)
	spira n. sp.?*, trilobites (USGS Loc. 2107).	
7	Limestone, mechanical, fine-grained lithofacies; *Lytospira* sp.	9.0
8	Limestone, mechanical, oölitic; *Finkelnburgia bridgei**, *Acontiodus* sp. *C**, *A.* sp. *D**, *Acanthodus* sp. *A**, *Distacodus* sp. *B**, *Drepanodus* sp. *A**, and *Paltodus* sp. *A** at base; *Lytospira* sp. (USGS Loc. 2108).	4.5
9	Limestone, mechanical, fine-grained lithofacies; a thin oölitic zone near base; *Tentaculites lowdoni** 21–26 and 32 feet above base, *Finkelnburgia* sp.*, *Acontiodus* sp. *A*, *B*, and *D**, *Acanthodus* sp. *A**, *Distacodus* sp. *A**, *Drepanodus* sp. *A* and *B**, *Loxodus bransoni**, *Paltodus* sp. *A**, *Paltodus?* sp., and *Scandodus* sp. 21–26 feet above base, *Lytospira* n. sp.* 42 feet above base, *Bellefontia* sp., pelmatozoan plates and columnals*, small, moderately high-spired gastropods (USGS Locs. 2109, 2110, 2111, 2112). (Traverse offset 15 feet north along strike.)	48.0
10	Limestone, mechanical, fine-grained lithofacies; *Tentaculites* sp. in lower foot.	14.0
	Measured thickness of Stonehenge limestone (incomplete, 5 per cent covered)	225.5
	Rockdale Run formation: At base is a 3-foot bed of laminated dolomitic limestone overlain by 30 feet of mechanical limestone followed by a covered interval containing cryptozöon chert to fence.	

Geologic Section 2—Stonehenge Limestone (Stoufferstown Member and Upper Member In Part)

Section begins at base of Stonehenge limestone in wooded area near northwest corner of Hoover farm 2.2 miles east of the center of Chambersburg and extends westward into pastures of Franklin County Home. Top of Stonehenge not exposed.

Unit	Description	Thickness (feet)
	Conochocheague formation: Thin-bedded fine-grained mechanical limestone without silty partings interbedded with laminated dolomitic limestone, pure limestone, and mottled dolomitic limestone; no fossils seen; about 100 feet poorly exposed to axis of anticline.	
	Stonehenge limestone Stoufferstown member	
1	Limestone, mechanical, coarse-grained lithofacies; top and bottom poorly exposed; about 50 per cent covered. (Traverse intersects fence about 90 feet above base.)	224.7
	Upper member	
2	Limestone, algal?, massive; small gastropods.	93.0
3	Limestone, mechanical, fine-grained lithofacies.	35.0
4	Covered.	99.6
5	Limestone, algal?, massive, about 50 per cent covered.	29.0
6	Limestone, mechanical, fine-grained lithofacies; gastropods; about 50 per cent covered.	25.0
7	Limestone, algal?, massive; gastropods and cephalopods.	17.0
8	Limestone, mechanical, fine-grained lithofacies.	4.0
9	Limestone, algal, channeled; gastropods.	23.0
10	Limestone, mechanical, fine-grained lithofacies; gastropods and cephalopods.	5.0
11	Limestone, algal?, massive; gastropods and cephalopods.	16.0
12	Limestone, mechanical, fine-grained lithofacies; brachiopods and gastropods.	19.0
13	Covered. (Traverse offset 21 feet south along strike.)	32.9
14	Limestone, mechanical, fine-grained lithofacies; brachiopods, gastropods, and pelmatozoan remains; about 50 per cent covered.	46.0
15	Covered. (Traverse offset 15 feet south along strike.)	56.2
16	Limestone, mechanical, fine-grained lithofacies; brachiopods, gastropods, and pelmatozoan remains; lower half about 50 per cent covered.	44.0
17	Dolomitic limestone, laminated.	3.0
18	Limestone, mechanical, interbedded coarse-grained (about 60 per cent) and fine-grained lithofacies; some oölitic beds containing asaphid trilobite fragments and gastropods; about 50 per cent covered. (Traverse intersects lane 17 feet above base and enters field west of pasture.)	144.0
	Measured thickness of upper member	691.7
	(Section ends in covered interval extending about 100 feet to base of ridge littered with cryptozöon chert. Top of Stonehenge limestone thought to occur at or near base of this ridge).	

Summary

Unit	Description	Thickness (feet)
	Measured thickness of upper member	691.7
	Stoufferstown member	224.7
	Measured thickness of Stonehenge limestone (incomplete, 47 per cent covered)	916.4

Geologic Section 3—Stonehenge Limestone (Stoufferstown Member and Upper Member In Part)

Section begins at south end of large pasture on Hoover farm about 70 feet northeast of a northwest-southeast fence at base of Stoufferstown member exposed at base of east flank of prominent ridge. Stoufferstown member and about lower half of upper member of Stonehenge limestone measured. Type section of Stoufferstown member.

Unit	Description	Thickness (feet)
	Conochocheague formation: Thin-bedded fine-grained mechanical limestone without silty partings interbedded with laminated dolomitic limestone, pure laminated limestone, and mottled dolomitic limestone; a zone of nodular black chert near base of exposures; no fossils seen; about 100 feet exposed to covered interval in which lies axis of major anticline.	
	Stonehenge limestone Stoufferstown member	
1	Limestone, mechanical, dominantly fine-grained lithofacies with a few calcirudite zones.	62.0
2	Limestone, oölitic, about 90 per cent concentric-structured oöliths about 0.75 mm. in diameter; *Pseudokainella*? aff. *P.*? *armatus**, *Hyolithes* sp.*, *Finkelnburgia* sp.*, ophiletid gastropod*, inarticulate brachiopod fragments*, merostome? fragments*, orthoceracone cephalopod*, *Aconticdus* sp.*, *Belodus* sp. *A**, *Cyrtoniodus*? sp.*, *Distacodus* cf. sp. *B**, and *Drepanodus* sp.*; some of the fossils collected north of traverse line (USGS Locs. 2113, 2131, 2132). (Traverse offset 75 feet south along strike, crossing fence at intersection with a northeast-striking fence and entering pasture to south.)	1.0
3	Limestone, mechanical, coarse-grained lithofacies. (Traverse offset 80 feet south along strike.)	48.0
4	Limestone, mechanical, coarse-grained lithofacies; indeterminate trilobite (possibly *Leiostegium*)* and indeterminate conodont* 5 feet above base; about 50 per cent covered; top poorly exposed (USGS Loc. 2114).	106.4
	Total Stoufferstown member	217.4
	Upper member	
5	Limestone, algal?, massive; poorly exposed; base just west of board fence on east side of lane leading to large pasture.	18.2
6	Covered; occupied by lane leading to large pasture. (Traverse enters pasture west of lane at top of unit.)	16.3
7	Limestone, algal?, massive; gastropods.	19.7
8	Limestone, mechanical, fine-grained lithofacies; gastropods; about 50 per cent covered. (Traverse offset 40 feet north along strike.)	75.4
9	Limestone, mechanical, fine-grained lithofacies.	13.8
10	Covered.	14.8
11	Limestone, mechanical, fine-grained lithofacies; poorly exposed.	17.7
12	Limestone, mechanical, fine-grained lithofacies.	19.7
13	Covered.	17.7
14	Limestone, mechanical, fine-grained lithofacies.	11.8
15	Limestone, algal, channeled; gymnosolenid stromatolites; gastropods. (Section ends at junction of three fences and gulley that strikes northwest; gulley marks trace of a fault.)	13.0
	Measured thickness of upper member	238.1

Summary

	Measured thickness of upper member	238.1
	Stoufferstown member	217.4
	Measured thickness of Stonehenge limestone (incomplete, 31 per cent covered)	455.5

Geologic Section 4—Stonehenge Limestone In Part and Rockdale Run Formation In Part

Section begins in beds near base of upper member of Stonehenge limestone exposed on east flank of hill in pasture on Hoover farm west of house and extends westward across Franklin County Home to farm west of Home. Traverse begins at base of hill about 150 feet west of small stone bridge on Falling Spring Branch. Northwest-striking gulley, representing the trace of a fault, separates this section from Geologic Section 3.

Stonehenge limestone
Upper member

Unit	Description	Thickness (feet)
1	Limestone, mechanical, fine-grained lithofacies; *Lytospira*? sp. 9 feet below top, trilobites, gastropods, and pelmatozoan debris throughout.	71.0
2	Covered.	26.3
3	Limestone, mechanical, fine-grained lithofacies.	23.0
4	Limestone, mechanical, coarse-grained lithofacies; *Lytospira*? sp., trilobites, pelmatozoan debris. (Traverse offset 50 feet south along strike.)	11.0
5	Limestone, mechanical, coarse-grained lithofacies; *Finkelnburgia bridgei** at top (USGS Loc. 2115). (Traverse offset 120 feet north along strike.)	20.0
6	Limestone, mechanical, coarse-grained lithofacies; trilobites and pelmatozoan debris.	53.0
7	Limestone, mechanical, fine-grained lithofacies; trilobites and pelmatozoan debris.	12.0
8	Limestone, mechanical, coarse-grained lithofacies; *Lytospira*? sp., trilobites, and pelmatozoan debris.	52.1
9	Limestone, mechanical, oölitic; *Lytospira* sp.*, *Bellefontia collieana*?*, merostome fragments*, and indeterminate conodonts*. (USGS Loc. 2116).	2.0
10	Limestone, mechanical, coarse-grained lithofacies; *Finkelnburgia* sp.*, inarticulate brachiopod fragments*, and *Drepanodus* sp. *A* and *D** 15.1 feet above base (USGS Loc. 2117).	24.4
11	Limestone, mechanical, fine-grained lithofacies; *Lytospira*? sp., trilobites, and gastropods.	16.5
12	Limestone, mechanical, coarse-grained lithofacies.	4.0
13	Limestone, mechanical, fine-grained lithofacies; *Lytospira*? sp.	20.9
14	Limestone, mechanical, coarse-grained lithofacies.	7.7
15	Limestone, mechanical, fine-grained lithofacies; *Lytospira*? sp. and trilobites.	8.4
16	Limestone, mechanical, coarse-grained lithofacies; *Lytospira*? sp.	7.7
17	Covered, but with a few outcrops of mechanical limestone; more complete exposures 200 feet north of traverse line show an oölitic bed 49 feet above base bearing *Lytospira*? sp., trilobites*, *Aconliodus* sp. *E**, *Acanthodus* sp. *A**, *Distacodus* sp. *A**, *Drepanodus* sp. *C* and *D**, *Oistodus* sp.*, and *Scolopodus* sp.* (USGS Loc. 2118). (Traverse intersects fence 60 feet above base and enters pasture of Franklin County Home.)	96.1
18	Limestone, mechanical, interbedded fine-grained and coarse-grained lithofacies, dominantly the former; oölitic in part; *Tentaculites* sp., gastropods, and brachiopods.	37.1
	Measured thickness of Stonehenge limestone (upper member).....	493.2

Rockdale Run formation

Unit	Description	Thickness (feet)
19	Dolomitic limestone, laminated. (Traverse offset 60 feet north along strike.)	1.0
20	Limestone, mechanical, fine-grained lithofacies; *Tentaculites lowdoni** and an indeterminate conodont* in upper 2 feet, pelmatozoan debris throughout (USGS Loc. 2119).	21.6
21	Covered; a few outcrops of mottled limestone and mechanical limestone near top.	136.3
22	Limestone, mechanical, calcilutite and calcisiltite, pure.	3.0
23	Limestone, algal; cryptozöon heads 7 inches in diameter and 3 inches high, consisting of broad, gentle upwarping of laminae.	1.0
24	Limestone, mechanical, calcilutite and calcisiltite, pure; nodular black chert parallels bedding. (Traverse offset 20 feet south along strike.)	4.0
25	Dolomitic limestone; laminated.	2.0
26	Limestone, mechanical, calcilutite and calcisiltite, pure; strongly sheared.	1.0
27	Covered.	131.0
28	Limestone, mechanical, dominantly fine-grained lithofacies; some dolomitic banding and mottling; *Lecanospira* sp.* and *Aconliodus* sp. *A** 31.5 feet above base (USGS Loc. 2120).	61.0
29	Limestone, mechanical, oölitic; oöliths 0.25–0.5 mm in diameter; *Lecanospira* sp. and *Macluritella* sp. (Good exposures of *Lecanospira* zone and beds immediately above and below it may be seen in abandoned quarry about 400 feet south of traverse line.) (Traverse offset about 650 feet north along strike, enters pasture to north; upper *Lecanospira* bed used as a datum.)	6.5
30	Covered; a few outcrops of mechanical limestone.	15.0
31	Dolomite; laminated.	2.0

Unit	Description	Thickness (feet)
32	Limestone, algal, massive; gymnosolenid stromatolites; small gastropods; about 50 per cent covered.	28.0
33	Covered; a few outcrops of mechanical limestone.	78.0
34	Limestone, mechanical, dominantly fine-grained lithofacies; endoceroid cephalopod siphuncles* 5 feet above base; *Leiostegium* cf. *L. puteatum**, macluritid gastropod*, high-spired gastropod*, pelmatozoan fragments*, and indeterminate conodonts* 7 feet above base (USGS Locs. 2121, 2122). (Traverse intersects fence 7 feet above base and enters pasture to west.)	60.0
35	Limestone, interbedded algal (about 60 per cent) and mechanical; stromatolites mainly gymnosolenid type but at least one zone of pycnostromids; gastropods; pelmatozoan debris, and cephalopods common throughout; *Escharendoceras simplex**, other endoceroid siphuncles*, indeterminate brachiopod*, and *Scolopodus quadraplicatus** 3 feet above base; *Archaeoscyphia annulata?** 36 feet above base (USGS Locs. 2123, 2124).	90.0
36	Covered; a few outcrops of mechanical and algal limestones; outcrops just above base of unit south of traverse line yielded *Diparelasma?* sp.*, *Mcqueenoceras?* sp.*, and indeterminate endoceroid siphuncles*, ophiletid gastropods*, and an indeterminate conodont* (USGS Locs. 2133, 2134, 2135).	70.5
37	Limestone, interbedded algal (about 50 per cent) and mechanical; gymnosolenid stromatolites; gastropods, pelmatozoa, and cephalopod siphuncles throughout.	23.5
38	Dolomitic limestone; mottled; appears to represent partial dolomitization of mechanical limestone.	1.5
39	Limestone, algal; gymnosolenid stromatolites.	9.0
40	Dolomite, laminated.	4.0
41	Limestone, interbedded algal (about 60 per cent) and mechanical; gymnosolenid stromatolites.	31.0
42	Orthoquartzite; well-rounded quartz sand grains 1 mm or less in diameter in a calcareous and dolomitic matrix along with a few pebbles of dolomite. (Traverse offset 150 feet south along strike.)	8.5
43	Limestone, interbedded algal (about 60 per cent) and mechanical; gymnosolenid stromatolites.	26.5
44	Dolomitic limestone; mottled, like unit 38.	1.5
45	Limestone, interbedded algal (about 60 per cent) and mechanical; gymnosolenid stromatolites.	31.0
46	Covered and unreliable outcrops of limestone.	69.0
47	Covered; interval obscured by cornfield; distance paced.	102.0
48	Covered; interval obscured by garden and weeds; a few outcrops of limestone; distance paced.	111.0
49	Covered; interval occupied by orchard and sewage disposal plant.	172.1
50	Limestone, mechanical, dominantly fine-grained lithofacies; some dolomitic mottling and banding; *Diparelasma* cf. *D. marylandicum**, *Tritoechia* sp.*, *Isoteloides* cf. *I. flexus**, ophiletid gastropod*, pleurotomarian gastropod*, *Acodus* sp.*, *Drepanodus* sp. *A* and *C**, and *Oistodus* sp. *A* and *B** 15 feet above base; gastropods and pelmatozoan debris abundant throughout (USGS Loc. 2125). (Traverse offset about 200 feet north along strike, entering pasture to north.)	76.0
51	Covered; interval occupied by dirt road and cow path.	28.9
52	Limestone, mechanical (about 70 per cent) and algal; gastropods, cephalopods, and pelmatozoan debris throughout; about 60 per cent covered. (Traverse intersects fence and enters field west of pasture.)	261.0
53	Covered; a few outcrops of limestone and dolomite give control to the traverse; blocky chert common in soil. (Traverse intersects fence at top of unit.)	658.0
54	Covered; interval occupied by road. (Traverse enters field west of road.)	32.5
55	Covered; distance scaled from aerial photo; section ends at point on axis of major syncline projected from position of outcrops to south.	188.0
	Measured thickness of Rockdale Run formation	2546.9

Summary

Measured thickness of Rockdale Run fortion (incomplete, 74 per cent covered)	2546.9
Measured thickness of Stonehenge limestone (upper member) (incomplete, 25 per cent covered)	493.7

Geologic Section 5—Rockdale Run Formation (In Part)

Traverse across quarry on north side of Falling Spring Branch just north of U. S. Highway 30, 1.3 mi. southeast of the center of Chambersburg. Traverse begins just opposite northeast corner of corn crib in southeast quarter of quarry; first bed measured along quarry rim.

Unit	Description	Thickness (feet)
1	Dolomite; laminated; upper 4 inches gradational into unit above.	1.3
2	Dolomitic limestone; mottled; gradational into unit above.	1.0
3	Dolomite; laminated.	3.0
4	Dolomitic limestone; mottled.	2.0
5	Dolomite; laminated.	0.8
6.	Dolomitic limestone; mottled; grades into unit above; leperditid ostracods* (USGS loc. 2183).	8.3
7	Dolomite; laminated.	2.2
8	Limestone; structureless; pure, massive.	1.5
9	Dolomite; laminated.	0.7
10	Dolomitic limestone; mottled.	2.0
11	Dolomitic limestone; laminated.	1.5
12	Limestone; structureless; thin-bedded, pure calcilutite; ostracods.	3.5
13	Dolomite; laminated.	4.0
14	Limestone; structureless; pure.	1.5
15	Dolomite; laminated.	3.5
16	Dolomitic limestone; mottled; dolomite in ribbon and bands in upper half; ostracods.	11.0
17	Dolomite; laminated.	0.8
18	Dolomitic limestone; laminated.	2.5
19	Dolomite; laminated.	2.0
20	Dolomitic limestone; mottled; probable fault.	9.0
21	Dolomite; laminated.	1.7
22	Limestone; structureless; pure.	1.5
23	Dolomite; laminated.	2.5
24	Dolomitic limestone; mottled.	6.0
25	Dolomitic limestone; laminated.	6.0
26	Dolomitic limestone; mottled.	4.0
27	Limestone; laminated.	1.7
28	Dolomite; laminated.	5.0
29	Dolomitic limestone; mottled; irregular white chert and calcite; rare gastropods and ostracods.	12.0
30	Dolomite; laminated.	5.0
31	Limestone; structureless; pure.	3.0
32	Dolomitic limestone; mottled.	2.0
33	Limestone; structureless; pure.	2.5
34	Limestone; laminated.	6.0
35	Limestone; structureless; pure.	1.5
36	Dolomitic limestone; laminated.	8.0
37	Dolomitic limestone; mottled.	4.0
38	Dolomite; laminated.	0.7
39	Dolomitic limestone; mottled.	3.7
40	Dolomite; laminated.	1.0
41	Dolomitic limestone; mottled.	9.0
42	Limestone; mechanical; pure calcilutite; ostracods.	4.0
43	Dolomitic limestone; laminated.	0.5
44	Dolomitic limestone; mottled.	2.0
45	Dolomitic limestone; laminated.	0.9
46	Dolomitic limestone; mottled; ostracods.	5.0
47	Dolomitic limestone; laminated.	1.0
48	Dolomitic limestone; mottled.	3.5
49	Dolomite; laminated.	5.0
50	Limestone; laminated; grades into unit below.	2.0
51	Dolomitic limestone; mottled; ostracods.	8.3
52	Dolomitic limestone; laminated.	0.7
53	Dolomitic limestone; mottled and banded.	7.7
54	Dolomitic limestone; laminated.	0.6
55	Dolomitic limestone; mottled and banded.	11.0
56	Dolomitic limestone; laminated.	4.0
57	Dolomite; laminated.	3.0
58	Dolomitic limestone; mottled.	9.5
59	Dolomite; laminated.	1.2
60	Dolomitic limestone; mottled; *Anomalorthis*? cf. *A. vermontensis**, *Clelandoceras* sp.*, ostracods, discoidal gastropods, and murchisonid snails (USGS Loc. 2184); this unit may be the interval from which USGS Loc. 291B-OS was collected.	12.0
61	Limestone; mechanical; fine-grained lithofacies; thin-bedded.	4.3
62	Dolomite; laminated.	1.5
63	Limestone; mechanical; like unit 60; a zone of dolomite pebbles at base.	3.4
64	Dolomitic limestone; laminated.	2.0
65	Dolomitic limestone; mottled.	11.0
66	Limestone; algal; pure; pycnostromid stromatolites.	2.0
67	Dolomite; laminated.	3.4
68	Dolomitic limestone; mottled and banded; thin-bedded.	6.5
69	Dolomite; laminated.	4.0
70	Dolomitic limestone; mottled; irregular white chert; discoidal gastropods* (USGS loc. 2185).	2.0
71	Dolomite; laminated.	4.5
72	Limestone; structureless; irregular white chert.	1.0
73	Limestone; laminated.	2.0
74	Dolomite; laminated.	6.0
75	Limestone; structureless.	1.5
76	Dolomitic limestone; mottled.	3.5
	Total Rockdale Run formation measured (incomplete)	289.4

Units 60–75 were measured in a ravine that enters west side of quarry. Section ends at a culvert in ravine.

Thickness (feet)
9.0
4.0
0.5
2.0
0.9
5.0
1.0
3.5
5.0
2.0
8.3
0.7
7.7
0.6
11.0
4.0
3.0
9.5
1.2
12.0
4.3
1.5
3.4
2.0
11.0
2.0
3.4
6.5
4.0
2.0
4.5
1.0
2.0
6.0
1.5
3.5
289.4

FOSSIL LOCALITIES EXCLUSIVE OF MEASURED SECTIONS

The following is a list of important U. S. Geological Survey localities in the Chambersburg area from which collections of fossils were identified by the writer. All localities are shown on Figure 2. Collections designated OS were made by E. O. Ulrich, G. W. Stose, and H. E. Dickhaut between 1906 and 1908. Specific names used by Ulrich (*in* Stose, 1909, p. 7) are in parentheses following the writer's identifications. The stratigraphic positions of these collections were interpreted from original locality data and faunal content. Other collections were made by the writer in 1955.

290Y-OS. This collection was made from a stone fence; clearly two faunal zones are represented:

(1) Rockdale Run formation, *Lecanospira* zone:
Lecanospira sp.

(2) Rockdale Run formation, *Diparelasma* zone (*Isoteloides* subzone):
Tritoechia sp. (*Orthis* cf. *O. electra*)
Rhombella sp. (*Liospira canadensis*)
Planispiral gastropod, indet. (*Maclurea affinis*)
High-spired gastropod, indet.
Isoteloides cf. *I. flexus* (*Isotelus canalis*)
Goniotelina sp. (*Bathyurus caudatus*)
Pseudocybele? sp. (*Amphion* cf. *A. salteri*)
Dimeropygiella? sp.
Trilobite, indet.

290Z-OS. The fossils in this collection suggest a zone of earliest Middle Ordovician age (Whiterock stage of Cooper, 1956). Inasmuch as the collection was made from a stone fence, its position in the Ordovician sequence at Chambersburg is unknown.
Maclurites magnus?
Macluritid gastropod cf. *Palliseria*?

291-OS. Rockdale Run formation, *Archaeoscyphia* zone, approximately 750 feet above the base of the formation.
Brachiopod, indet.
Mysticoceras? cf. *M. fusiforme*
Platysiphon sp.
Piloceroid cephalopod, indet.
Endoceroid cephalopods, indet.

291A-OS. Rockdale Run formation, *Lecanospira* zone, exact stratigraphic position unknown.
Lecanospira sp.
Tarphyceracone cephalopod, indet.
Orthoceracone cephalopod, indet.

291B-OS. Rockdale Run formation, *Syntrophopsis-Clelandoceras* zone, between approximately 1900 and 2200 feet above the base of the formation. Collection may have been made from unit 60 of Geologic Section 5.
Nothorthis pennsylvanica [*Orthis* (?*Dalmanella*) *electra*]
Maclurites cf. *M. oceanus* (*Maclurea* cf. *M. oceana*)
Helicotoma? *disjuncta* (*Ophileta*? *disjuncta*)
Murchisonia (*Turritoma*) cf. *acrea* (*Turritoma acrea*?)
Moderately high-spired gastropod, indet. (*Lophospira* cf. *Murchisonia gregaria* Billings)
Murchisonid gastropod, indet. (*Hormotoma gracilens*)
Planispiral gastropod, indet.
Clelandoceras sp. (*Cyrtocerina* cf. *C. mercurius*)
Curtoceras internastriatum (*Trocholites internestriatus*)
Orthoceracone cephalopod, indet.
Eoleperditia sp.

291C-OS. Rockdale Run formation, *Syntrophopsis-Clelandoceras* zone, approximately 1700 feet above the base of the formation.
Syntrophopsis? sp. (*Syntrophia lateralis*)
Brachiopod, indet.
Maclurites sp. (*Maclurea*? *sordida*)
Liospirid gastropod, indet. (*Liospira* cf. *L. laurentia*)
Orthoceracone cephalopod, indet. (*Orthoceras* cf. *O. primigenium*)

291C1-OS. Rockdale Run formation, *Syntrophopsis-Clelandoceras* zone, between approximately 1900 and 2200 feet above the base of the formation. Collection may have been made from Unit 60 of Geologic Section 5.
Close-coiled, possibly hyperstrophic gastropod, indet.

291C2-OS. Rockdale Run formation, *Syntrophopsis-Clelandoceras* zone, between approximately 1900 and 2200 feet above the base of the formation. Collection may have been made from Unit 60 of Geologic Section 5.
Murchisonia (*Hormotoma*?) cf. *H. confusa*
Clelandoceras sp.

291C3-OS. Rockdale Run formation, *Syntrophopsis-Clelandoceras* zone, between approximately 1900 and 2200 feet above the base of the formation. Collection may have been made from Unit 60 of Geologic Section 5.
Ectomaria sp. (*Solenospira* cf. *S. prisca*)
Moderately high-spired gastropod, indet.

292Q-OS. Stonehenge limestone, near the base of the upper member.
Ophiletid gastropod, indet. (*Ophileta complanata*)
Hystricurid trilobite, indet.

2126. Stonehenge limestone, about 100 feet below top.
Finkelnburgia stonehengensis

2127. Rockdale Run formation, *Diparelasma* zone (*Isoteloides* subzone), about 1200 feet above the base of the formation.
Diparelasma sp.
Inarticulate brachiopod fragments, indet.
Gastropods, indet.
Isoteloides cf. *I. flexus*

Isochilina sp. (*Leperditia* cf. *Primitia gregaria* Whitfield)
Acontiodus sp.
Acanthodus sp. *A*
Drepanodus sp. *A*
Oistodus sp. *A*
Scolopodus sp.

2128. Rockdale Run formation, *Diparelasma* zone, about 1000 feet above the base of the formation.

Tritoechia sp.

2129. Rockdale Run formation, *Diparelasma* zone (*Isoteloides* subzone), about 1200 feet above the base of the formation.

Pelmatozoan fragments, indet.
Tritoechia sp.
Gastropods, indet.
Endoceroid cephalopods, indet.
Isoteloides cf. *I. flexus*
Acontiodus sp. *A*
Scolopodus quadraplicatus

2130. Rockdale Run formation, *Lecanospira* zone, about 500 feet above the base of the formation.

Finkelnburgia sp.

References Cited

Bass, N. W., and Northrop, S. A., 1953, Dotsero and Manitou formations, White River Plateau, Colorado, with special reference to Clinetop algal limestone member of Dotsero formation: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 889–912

Bassler, R. S., 1919, Report on the Cambrian and Ordovician formations of Maryland: Md. Geol. Survey Special Pub., 424 p.

Brainerd, Ezra, and Seely, H. M., 1890, The Calciferous formation in the Champlain Valley: Geol. Soc. America Bull., v. 1, p. 501–516

Branson, E. B., and Mehl, M. G., 1933, Conodonts from the Jefferson City (Lower Ordovician) of Missouri: Mo. Univ. Studies, v. 8, p. 53–64

Bridge, Josiah, and Cloud, P. E., Jr., 1947, New gastropods and trilobites critical in the correlation of Lower Ordovician rocks: Am. Jour. Sci., v. 245, p. 545–559

Cady, W. M., 1945, Stratigraphy and structure of west-central Vermont: Geol. Soc. America Bull., v. 56, p. 515–558

Cooper, G. A., 1956, Chazyan and related brachiopods: Smithson. Misc. Coll., v. 127, 1024 p.

Cullison, J. S., 1944, The stratigraphy of some Lower Ordovician formations of the Ozark Uplift: Mo. Univ. School Mines and Metallurgy Tech. Ser. Bull., v. 15, no. 2, 112 p.

Flower, R. H., 1956, Cephalopods from the Canadian of Maryland: Jour. Paleontology, v. 30, p. 75–96

Furnish, W. M., 1938, Conodonts from the Prairie du Chien beds of the Upper Mississippi Valley: Jour. Paleontology, v. 12, p. 318–340

Hintze, L. F., 1952, Lower Ordovician trilobites from western Utah and eastern Nevada: Utah Geol. Mineralog. Survey Bull. 48, 249 p.

Lochman, Christina, and Duncan, D. C., 1950, The Lower Ordovician *Bellefontia* fauna in central Montana: Jour. Paleontology, v. 24, p. 350–353

Neuman, R. B., 1951, St. Paul group: a revision of the "Stones River" group of Maryland and adjacent states: Geol. Soc. America Bull., v. 62, p. 267–324

Nolan, T. B., Merriam, C. W., and Williams, J. S., 1956, The stratigraphic section in the vicinity of Eureka, Nevada: U. S. Geol. Survey Prof. Paper 276, 77 p.

Raymond, P. E., 1924, New Upper Cambrian and Lower Ordovician trilobites from Vermont: Boston Soc. Nat. History Proc., v. 37, p. 389–466

—— 1937, Upper Cambrian and Lower Ordovician Trilobita and Ostracoda from Vermont: Geol. Soc. America Bull., v. 48, p. 1079–1146

Ross, R. J., Jr., 1951, Stratigraphy of the Garden City formation in northeastern Utah, and its trilobite faunas: Peabody Mus. Nat. History Bull. 6, 161 p.

Sando, W. J., 1956, Nomenclature of Lower Ordovician rocks of Washington County, Maryland: Geol. Soc. America Bull., v. 67, p. 935–938

—— 1957, Beekmantown group (Lower Ordovician) of Maryland: Geol. Soc. America Mem. 68, 161 p.

Stose, G. W., 1908, The Cambro-Ordovician limestones of the Appalachian Valley in southern Pennsylvania: Jour. Geology, v. 16, p. 698–714

—— 1909, Description of Mercersburg-Chambersburg district: U. S. Geol. Survey Geol. Atlas, Folio 170, 19 p.

Twenhofel, W. H., *et al.*, 1954, Correlation of the Ordovician formations of North America: Geol. Soc. America Bull., v. 65, p. 247–298

Ulrich, E. O., 1911, Revision of the Paleozoic systems: Geol. Soc. America Bull., v. 22, p. 281–680

Ulrich, E. O., and Cooper, G. A., 1938, Ozarkian and Canadian Brachiopoda: Geol. Soc. America Special Paper 13, 323 p.

Ulrich, E. O., Foerste, A. F., Miller, A. K., and Furnish, W. M., 1942, Ozarkian and Canadian cephalopods, Part 1: Nautilicones: Geol. Soc. America Special Paper 37, 157 p.

Ulrich, E. O., Foerste, A. F., Miller, A. K., and Unklesbay, A. G., 1944, Ozarkian and Canadian cephalopods, Part 3: Longicones and summary: Geol. Soc. America Special Paper 58, 226 p.

Unklesbay, A. G., and Young, R. S., 1956, Early Ordovician nautiloids from Virginia: Jour. Paleontology, v. 30, p. 481–491

Walcott, C. D., 1924, Cambrian geology and paleontology V; No. 1, Geological formations of Beaverfoot-Brisco-Stanford Range, British Columbia, Canada: Smithson. Misc. Coll., v. 75, no. 1, p. 1–51

—— 1925, Cambrian geology and paleontology V; No. 3, Cambrian and Ozarkian trilobites: Smithson. Misc. Coll., v. 75, no. 3, p. 61–146

Whitfield, R. P., 1890, Observations on the fauna of the rocks at Fort Cassin, Vermont, with descriptions of a few new species: Am. Mus. Nat. History, v. 3, p. 25–39

Whittington, H. B., 1953, North American Bathyuridae and Leiostegiidae (Trilobita): Jour. Paleontology, v. 27, p. 647–678

Wilson, J. L., 1952, Upper Cambrian stratigraphy in the central Appalachians: Geol. Soc. America Bull., v. 63, p. 275–322

Yochelson, E. L., and Bridge, Josiah, 1957, The Lower Ordovician gastropod *Ceratopea*: U. S. Geol. Survey Prof. Paper 294-H, p. 281–304

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Plate 2.—Lower Ordovician fossils

1. *Murchisonia* (*Turritoma*) cf. *M. acrea* Billings. Side view of latex cast, ×1, USNM 133394; Rockdale Run formation (*Syntrophopsis-Clelandoceras* zone), 1900–2000 feet above base of formation, USGS Loc. 291B-OS.
2. *Ectomaria* sp. Side view of latex cast, ×1, USNM 133395; Rockdale Run formation (*Syntrophopsis-Clelandoceras* zone), 1900–2000 feet above base of formation, USGS Loc. 291C3-OS.

3, 4. *Rhombella* sp. Side and dorsal views, respectively, of internal cast, ×1, USNM 133396; Rockdale Run formation (*Diparelasma* zone), 1200–1300 feet above base of formation, USGS Loc. 290Y-OS.

5–7. *Lytospira* n. sp. Dorsal, ventral, and apertural views, respectively, ×1, USNM 133397; Stonehenge limestone, 20 feet below top, USGS Loc. 2111.

8. *Helicotoma? disjuncta* (Billings). Dorsal view of latex cast, ×1, USNM 133398; Rockdale Run formation (*Syntrophopsis-Clelandoceras* zone), 1900–2000 feet above base of formation, USGS Loc. 291B-OS.
9. *Eoleperditia* sp. Side view of right valve, ×3, USNM 133399; Rockdale Run formation (*Syntrophopsis-Clelandoceras* zone), 1900–2000 feet above base of formation, USGS Loc. 291B-OS.
10. *Isochilina* sp. Side view of right valve, ×10, USNM 133400; Rockdale Run formation (*Diparelasma* zone), 1200 feet above base of formation, USGS Loc. 2127.

11–14. *Anomalorthis?* cf. *A. vermontensis* Ulrich and Cooper. Interior and exterior views of pedicle and brachial valves, respectively, ×2, USNM 133401 *a*, *b*; Rockdale Run formation (*Syntrophopsis-Clelandoceras* zone), 2130 feet above base of formation, USGS Loc. 2184.

15. *Hyolithes* sp. Side view, ×5, USNM 133402; Stonehenge limestone (Stoufferstown member), 62 feet above base of formation, USGS Loc. 2132.
16. *Tentaculites lowdoni* Fisher and Young. Side view, ×5, USNM 133403; Stonehenge limestone (upper member), 30 feet below top of formation, USGS Loc. 2110.
17. *Loxodus bransoni* Furnish. Inner side view, ×30, USNM 133404; Stonehenge limestone (upper member), 36–41 feet below top of formation, USGS Loc. 2109 (Figures 17–27 slightly retouched).
18. *Paltodus* sp. *A*. Inner side view, ×30, USNM 133405; Stonehenge limestone (upper member), 70 feet below top of formation, USGS Loc. 2108.
19. *Oistodus* sp. *A*. Outer side view, ×30, USNM 133406; Rockdale Run formation (*Diparelasma* zone), 1310 feet above base of formation, USGS Loc. 2125.
20. *Acanthodus* sp. *A*. Inner side view, ×30, USNM 133407; Stonehenge limestone (upper member), 36–41 feet below top of formation, USGS Loc. 2109.
21. *Scolopodus quadraplicatus* Branson and Mehl. Inner side view, ×30, USNM 133048; Rockdale Run formation (*Archaeoscyphia* zone), 550 feet above base of formation, USGS Loc. 2123.
22. *Distacodus* sp. *B*. Inner side view, ×30, USNM 133409; Stonehenge limestone (upper member), 140 feet below top of formation, USGS Loc. 2105.
23. *Belodus* sp. *A*. Outer side view, ×30, USNM 133410; Stonehenge limestone (Stoufferstown member), 62 feet above base of formation, USGS Loc. 2132.
24. *Acodus* sp. Inner side view, ×30, USNM 133411; Rockdale Run formation (*Diparelasma* zone), 1310 feet above base of formation, USGS Loc. 2125.
25. *Acontiodus* sp. *A*. Inner side view, ×30, USNM 133412; Stonehenge limestone (upper member), 36–41 feet below top of formation, USGS Loc. 2109.
26. *Cyrtoniodus?* sp. Outer side view, ×30, USNM 133413; Stonehenge limestone (Stoufferstown member), 62 feet above base of formation, USGS Loc. 2132.
27. *Drepanodus* sp. *A*. Inner side view, ×30, USNM 133414; Stonehenge limestone (upper member), 70 feet below top of formation, USGS Loc. 2108.

28, 29. *Leiostegium* cf. *L. puteatum* Raymond. Dorsal views of cranidium and pygidium, ×5 and ×2, respectively, USNM 133415 *a*, *b*; Rockdale Run formation (*Lecanospira* zone), 500 feet above base of formation, USGS Loc. 2121.

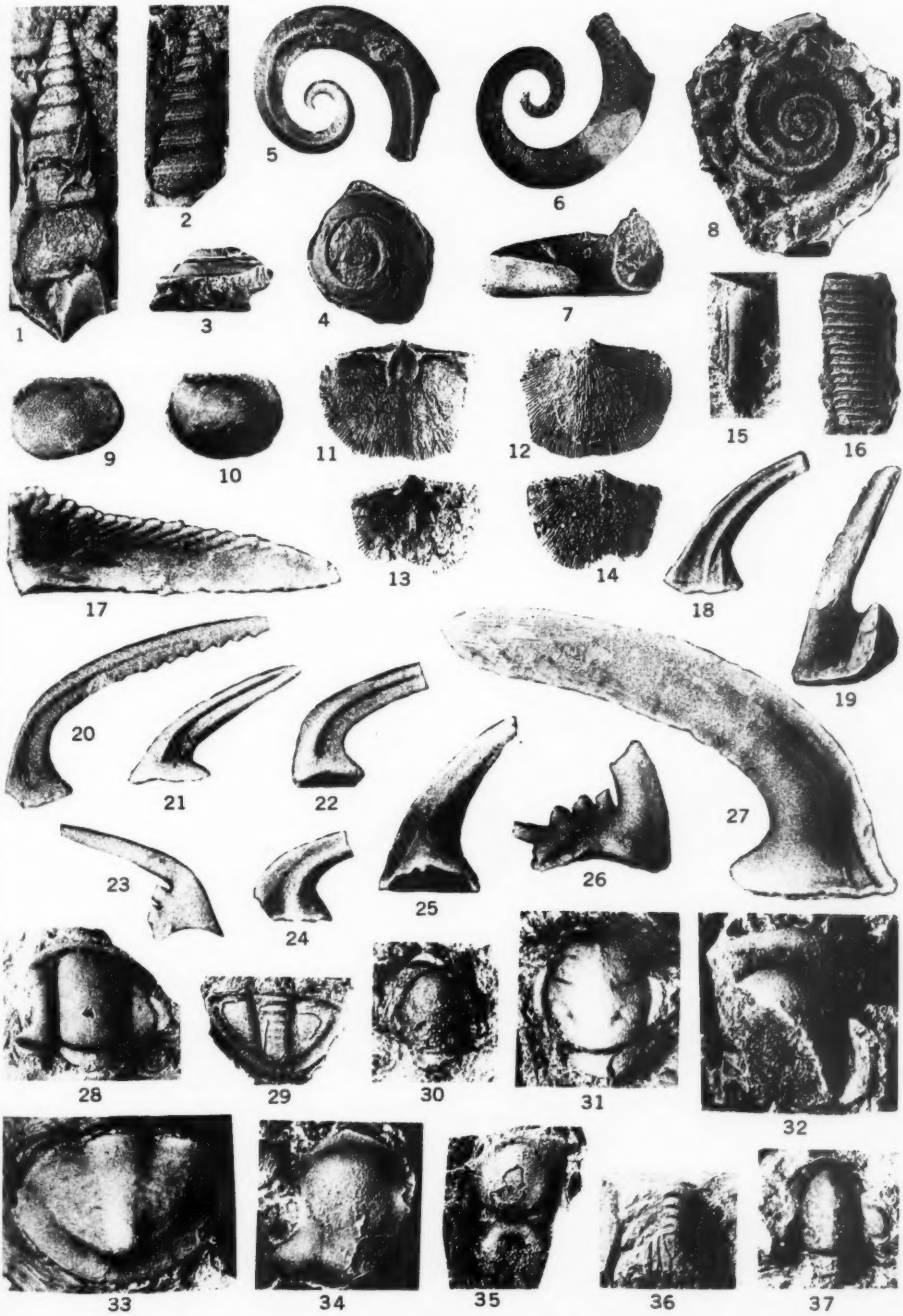
30. *Dimeropygiella?* sp. Dorsal view of cranidium, ×5, USNM 133416; Rockdale Run formation (*Diparelasma* zone), 1200–1300 feet above base of formation, USGS Loc. 290Y-OS.

31, 32. *Pseudokainella?* aff. *P.? armatus* Hintze. Dorsal views of two imperfect cranidia, ×4 and ×10, respectively, USNM 133417 *a*, *b*; Stonehenge limestone (Stoufferstown member), 62 feet above base of formation, USGS Loc. 2132.

33–35. *Isoteloides* cf. *I. flexus* Hintze. Dorsal views of pygidium and cranidium, ×4, and ventral view of hypostome, ×2, USNM 133418 *a–c*; Rockdale Run formation (*Diparelasma* zone), 1310 feet above base of formation, USGS Locs. 2125 and 292X-OS.

36. *Pseudocybele?* sp. Dorsal view of pygidium, ×5, USNM 133419; Rockdale Run formation (*Diparelasma* zone), 1200–1300 feet above base of formation, USGS Loc. 290Y-OS.

37. *Goniotelina* sp. Dorsal view of cranidium, ×3, USNM 133420; Rockdale Run formation (*Diparelasma* zone), 1200–1300 feet above base of formation, USGS Loc. 290Y-OS.



LOWER ORDOVICIAN FOSSILS

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METAMORPHOSED MIDDLE PALEOZOIC FOSSILS FROM CENTRAL MASSACHUSETTS, EASTERN VERMONT, AND WESTERN NEW HAMPSHIRE

BY ARTHUR J. BOUCOT, GORDON J. F. MACDONALD, CHARLES MILTON, AND JAMES B. THOMPSON, JR.

ABSTRACT

Study of thin and polished sections and spectrographic analyses indicate that the brachiopod most recently used to date the Bernardston Formation in Massachusetts probably came from Lower Devonian beds (chlorite zone) in Nova Scotia, and not from Bernardston, Massachusetts. Restudy of the faunule from the calcareous quartzite (garnet zone) of the Bernardston Formation indicates that it is probably of Silurian rather than Devonian age. The upper part of the Clough Formation on Skitchewaug Mountain (garnet zone) in eastern Vermont contains tetracorals of Silurian or Devonian age. The Clough Formation on Croydon Mountain (sillimanite zone) in west-central New Hampshire contains tetracorals. The fossiliferous zone in the lower part of the Bernardston Formation is possibly equivalent to the upper part of the Clough Formation.

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INTRODUCTION

In the metamorphosed strata of New England, fossils are exceedingly rare, and many of the specimens are deformed. Nevertheless, the fossils that have been found are important for the interpretation of the stratigraphy and the structure of the region. This paper re-examines carefully the fossils obtained from the metamorphosed rocks of the Connecticut Valley in Massachusetts, southern Vermont, and west-central New Hampshire and relates them to the stratigraphic succession. Boucot is responsible for the study of the fossils and the stratigraphic correlations; Richard S. Harner, formerly of the U. S. Geological Survey, studied the samples of magnetite spectrographically; MacDonald is responsible for the work on the stratigraphy of the Bernardston, Massachusetts, area; Thompson is responsible for the stratigraphy of the Skitchewaug Mountain and Croydon Mountain areas; and Milton made the

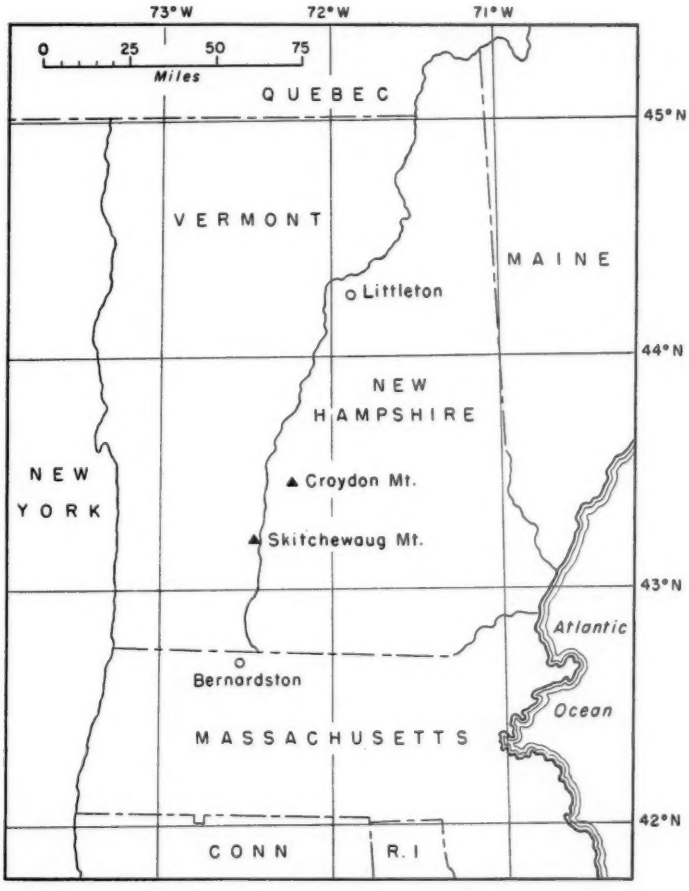

FIGURE 1.—LOCATION OF FOSSIL LOCALITIES MENTIONED IN TEXT

petrographic studies bearing on the provenance of a specimen whose locality was in dispute.

The aid of Dr. G. A. Cooper in the study of the brachiopods is gratefully acknowledged. The work of Professors E. C. Stumm (University of Michigan), W. H. Easton (University of Southern California), and Miss Helen Duncan of the U. S. Geological Survey on the tetracorals leaves us in their debt. We are also indebted to the Peabody Museum, Yale University, and to the Museum of Amherst College for the loan of collections. We have greatly profited from discussions with Professor Marland P. Billings, Harvard University, of the problems involved in dating the Bernardston Formation.

STRATIGRAPHY

The middle Paleozoic stratigraphy of the Connecticut Valley region is best known in the Littleton-Moosilauke area (Billings, 1937) of northwestern New Hampshire. The lowest unit of middle Paleozoic age in this area, unconformably overlying rocks that predate the Taconic orogeny, is the Clough Formation, which consists largely of quartzite and conglomerate. The overlying Fitch Formation, of Silurian age, consists largely of medium- to fine-grained clastic and argillitic material and is characterized by a high carbonate content. The Fitch Formation is succeeded by the medium- to fine-grained clastic and argillitic

material of the Littleton Formation, in whose middle portion occur fossils of late Early Devonian age.

South of the Littleton-Moosilauke area metamorphosed fossils of middle Paleozoic age have been found at three localities: Skitchewaug Mountain, east-central Vermont; Croydon Mountain, west-central New Hampshire; and near Bernardston, north-central Massachusetts. The stratigraphic relations and precise age of the beds at these localities are of great importance to the understanding of the regional geology (Fig. 1).

The fossil locality at Skitchewaug Mountain is in quartzite assigned by Thompson (1954) to the upper part of the Clough Formation. The Croydon Mountain locality appears to be at the same stratigraphic position, but neither faunule can be accurately dated. The stratigraphic position of the calcareous quartzite and limestone of the Bernardston Formation at Bernardston, on the other hand, is uncertain, but the faunule can be dated with a higher degree of reliability than can that at either of the other localities. Various interpretations of the stratigraphic position of the fossils obtained from the Bernardston Formation have been made.

Emerson (1898, p. 262–271) believed that the section exposed at what was then the Williams Farm—about a quarter of a mile northwest of Bernardston—was a normal one, with the succession, in ascending order, limestone, magnetite, calcareous quartzite, quartzite and conglomerate, and phyllite. He (p. 285) stated that conglomerate and quartzite also normally underlie the limestone at localities near by. Emerson concluded that the calcareous quartzite at the Williams Farm is younger than the underlying limestone. If quartzite and conglomerate are normally present at the base of the unit, Emerson's Bernardston Formation is probably equivalent to all the Silurian and Devonian units of adjacent New Hampshire (Clough, Fitch, and Littleton formations).

Bain (1936, p. 517) suggested that the conglomerate, quartzite, calcareous quartzite, and limestone at the Williams Farm are overturned, and that the calcareous quartzite, and the quartzite and conglomerate are older than the limestone. Bain's interpretation is plausible since the sequence at the Williams Farm consisting of limestone, magnetite, calcareous quartzite, and quartzite and conglomerate is reminiscent of the Clough and Fitch relationship in New Hampshire, but inverted.

If the sequence is inverted, the quartzites and conglomerate at the Williams Farm could very well be the same as the quartzites and conglomerate found by Emerson (1898, p. 285) near by and below limestone beds.

Balk (1956) indicates that the quartzites and conglomerate of the Bernardston Formation are near the base of that unit. He dates the Bernardston Formation as "Devonian?", presumably in accord with previously published accounts of its age.

Age of the Bernardston Formation

General Statement

The fossil locality at Bernardston was first described by Edward Hitchcock (1833, p. 295) and has been the subject of several investigations, notably by Dana (1873; 1877), C. H. Hitchcock (1877, p. 428–457), Whitfield (1883), Emerson (1898, p. 253–299), and Balk (1941; 1956). The most recent determination of the age of the Bernardston Formation is that of Cooper *et al.* (1942, p. 1747). On the basis of "... a fairly well preserved *Spirifer* ..." from the "... magnetite bed ...", submitted for study by the late Robert Balk, Cooper concluded that there was "... evidence for an Onondaga age. ..." The late Professor Percy Raymond suggested to Boucot that the provenance of the "*Spirifer*" from the magnetite bed was questionable because his own experience indicated that the magnetite bed was unfossiliferous. Through the courtesy of Professor G. W. Bain, Boucot was able to examine the internal impression of the "*Spirifer*" (Pl. 1, figs. 4–5), which belongs to the collection of Amherst College, and to secure a piece of the magnetite composing it. This fossil was very similar, both in its morphology and in lithology of the magnetite, to specimens from the Lower Devonian iron ores of Nova Scotia and was distinctly different lithologically from magnetite collected at Bernardston by Boucot. In order to test this possibility the Amherst College specimen was subjected to spectrographic and petrographic examination, as well as morphologic study, and was compared with magnetite specimens known to have originated in Nova Scotia and in Bernardston, Massachusetts.

Morphology of the "Spirifer"

The specimen in the collections of Amherst College is an internal impression of a brachial valve preserved in magnetite. It has been somewhat sheared, as evidenced by the disposition of the lateral plications (Pl. 1, figs. 4–5) relative to the fold.

BRACHIAL INTERIOR: The cardinalia consist of discrete hinge plates supported by short crural plates. The posterior end of the notothyrial cavity is occupied by the typical striated surface of the structure homologous in spiriferoids to a cardinal process. The fold is strongly costate, as are the flanks. The costae are angular in cross section and do not appear to bifurcate near the anterior margin. The specimen is elliptical in outline. The hinge line is straight.

COMPARISON: The Amherst College specimen can best be compared with the following: *Fimbrispirifer divaricatus* (Hall, 1857) (Pl. 1, fig. 10), *"Spirifer" bischofi* Roemer, 1858 (Pl. 1, figs. 1–3), and an undescribed genus and species of spiriferoid known only from the Lower Devonian of Nova Scotia (Pl. 1, figs. 6–9).

F. divaricatus differs in the nature of its ornamentation, which consists of relatively low, anteriorly dividing plications with a cross section more nearly round than are those of the Amherst College specimen. The genus *Fimbrispirifer* has crural plates (Pl. 1, fig. 11), as does the specimen in question, but the ornamentation precludes the assignment of the specimen to that genus.

"Spirifer" bischofi has ornamentation very similar to that of the Amherst College specimen, but lacks the crural plates (Pl. 1, fig. 1). The absence of crural plates indicates the affinities of *"S." bischofi* with the spiriferoids of the *Hysterolites* group rather than with those of the *Howellella* group, which includes *Fimbrispirifer* as well as the specimen in question.

The undescribed genus and species known only from the Lower Devonian of Nova Scotia, on the other hand, has ornamentation and cardinalia identical with those of the specimen in the Amherst College collection as far as can be determined from the available material. The morphology of the *"Spirifer"*, therefore, is consistent with the thesis that the specimen came from Nova Scotia, though of course it does not preclude other possibilities.

Petrography of the Magnetite and Associated Rock

For the mineralogical and chemical study, seven specimens were examined: four from the magnetite bed at Bernardston, Massachusetts, two from Nictaux, Nova Scotia, and magnetite from the critical fossil. These seven specimens were labeled 1 to 7 with nothing to indicate their origin to the petrographer or the spectrographer. Plate 5, thin sections of material from Bernardston (2 and 4), from Nictaux (6), and from the *"Spirifer"* from the Bernardston Formation (7), shows clearly the likeness of the two last, and their unlikeness to the two former. From the Bernardston Formation we have relatively high-grade metamorphic rock, garnet-biotite schist with coarsely crystallized magnetite; from the Nictaux area and in the *"Spirifer"* from the Bernardston Formation, low-grade chlorite carbonate siltstone with finer-grained magnetite.

Plate 6 (polished sections) clearly shows the relatively coarse crystallization of the magnetite from Bernardston (specimens 1, 3, 4) compared to the fine-grained ore from Nictaux, Nova Scotia (specimens 5, 6).

Spectrographic Analysis of the Magnetite

Magnetite from the Amherst College specimen, two samples of magnetite from Nictaux, Nova Scotia, and three samples from Bernardston (two of which were taken from the same block of magnetite) were examined spectrographically (Table 1).

Fourteen metals were determined quantitatively in these magnetites. Of these, six (V, Ca, Sr, Sc, Y, Be) agree with the division of the specimens on the basis of provenance (six of the seven samples were analyzed). One of them, No. 2, had little or no magnetite, but it was studied petrographically and belongs with the other Bernardston samples on the basis of its mineralogy and texture. The six metals are either absent from the magnetite from Massachusetts or are much less abundant.

For Mg, Ni, Cu, and perhaps Al, the evidence is also corroborative, though less striking. The magnetite from Nova Scotia contains more Mg and Ni, and as much or more Al, than the rocks from Massachusetts; Cu is consistently lower in the magnetite from Nova Scotia.

For the four metals remaining, Ti, Cr, Mn, and Ba, the evidence is ambiguous, and Mn

apparently contradicts the preceding data. None of these metals is concentrated in the magnetites of either group. The spectrographic results strongly suggest that the Amherst College specimen is allied to the material from Nictaux, Nova Scotia, rather than to that from Bernardston.

There is no assurance that these metals are actually part of the magnetite crystals. They may be in contaminating minerals, or perhaps inclusions of the matrix in the magnetite. Any contaminant visible under a binocular (×30 magnification) was removed before the analysis.

Ca, Mg, Al, Ti, and Mn appear to be relatively abundant (>0.1%) in the magnetite. In most of the 11 analyses of magnetite listed by Palache, Berman, and Frondel (1945, p. 770, 701), some or all of the elements Mg, Al, Ti, and Mn are present.

Minguzzi (1950, p. 119–144) examined four samples of magnetite from Elba, in which he found Mn, Ti, Al, and Mg, and smaller amounts of Si, Cu, Sn, V, Ca, and Ni. Except for Sn, which was not found in the magnetite from America, all these metals are also present in all the magnetite samples from Massachusetts, except No. 1, in which Ti was not detected. Silicon was not determined in the magnetite from Massachusetts, as quartz was used to dilute the samples before arcing.

Magnetite No. 1, from Bernardston, appears to be of extreme purity; the total constituents other than iron total only 0.2775 per cent; no single metal exceeds 0.1 per cent. Most of the magnetite appears to contain less admixture than that from Nova Scotia, as might be expected in view of its relative coarseness; this facilitates separation of magnetite from the gangue.

X-ray examination showed no significant variation in unit-cell size.

TABLE 1.—QUANTITATIVE SPECTROGRAPHIC ANALYSES OF MAGNETITE SAMPLES

	1	3†	4†	5	6	7
V	0.007	0.014	0.010	0.066	0.064	0.110
Ca	0.06	0.19	0.36	1.20	1.50	1.50
Sr*	0.0000	0.0000	0.0000	0.0056	0.0065	0.0060
Sc	0.0000	0.0000	0.0000	0.0005	0.0012	0.0004
Y	0.000	0.000	0.000	0.010	0.009	0.012
Be	0.0004	0.0000	0.0000	0.0024	0.0020	0.0010
Mg	0.024	0.110	0.100	0.240	0.180	0.180
Ni	0.0015	0.0030	0.0028	0.0100	0.0048	0.0052
Cu	0.0026	0.0034	0.0022	0.0020	0.0020	0.0015
Al	0.086	1.100	0.760	2.400	1.300	1.100
Ti	0.00	0.20	0.16	0.13	0.10	0.19
Cr	0.0000	0.0046	0.0026	0.0052	0.0030	0.0130
Mn	0.096	0.170	0.170	0.520	0.320	0.080
Ba	0.0000	0.0032	0.0019	0.0027	0.0010	0.0000

Looked for but not found: Ag, Au, Hg, Pt, Mo, W, Re, Ge, Sn, Pb, As, Sb, Bi, Zn, Cd, Tl, In, Co Ga, La, Ce, Th, Nb, Ta, U

1—107437 (USNM Min. Coll.) Bernardston, Mass.

3—107438 (USNM Min. Coll.) Bernardston, Mass.

4—107439 (USNM Min. Coll.) Bernardston, Mass.

5—125253 (USNM Fossil Coll.) Nictaux, N. S.

6—125252 (USNM Fossil Coll.) Nictaux, N. S.

7—Amherst College specimen

*0 in unit column means element not detected

† Two samples taken from the same specimen (a block about 1 cubic foot)

Descriptions of Fossils from the Bernardston Formation

As suggested to Boucot by the late Professor Percy Raymond, no fossils have been reported from the magnetite bed of the Bernardston Formation in any of the following investigations of the locality: E. Hitchcock (1833; 1852); Hitchcock and Hitchcock (1861); Dana (1873; 1877); C. H. Hitchcock (1877); Whitfield (1883); Emerson (1898); Clarke (1909). The late Robert Balk's abstract (1941) does mention fossils associated with the "magnetite bed." Fossils have been recorded from both the limestone (actually a marble) and the calcareous quartzite of the Bernardston Formation, and these fossils are redescribed here.

The catalog numbers of specimens belonging to the Peabody Museum of Natural History, Yale University, are preceded by Y.P.M., those of the U. S. National Museum by U.S.N.M.

Unidentified Tetracoral

(Pl. 3, figs. 2, 3)

Remarks.—Boucot collected a few small tetracorals from the calcareous quartzite bed of the Bernardston Formation. However, they are too poorly preserved to be generically identified. Simi-

lar material was evidently available to Whitfield (1883, p. 368) (his "*Petraia* or *Streptelasma*").

Occurrence.—Williams Farm, Bernardston, Mass.

Geologic location.—Calcareous quartzite of the Bernardston Formation.

Figured specimens.—U.S.N.M. 125264, 125263

Unidentified Favositid

(Pl. 3, fig. 1)

Remarks.—Tabulate corals are rare in the limestone of the Bernardston Formation, and Boucot found no specimens at the locality. One specimen in the collections of the Peabody Museum of Natural History, however, was collected at Bernardston and has the morphology, including closely spaced tabulae and large cells, of a favositid. It is too poorly preserved to be assigned generically. Similar material was collected by the late Professor Percy Raymond and is now in the collections of the Museum of Comparative Zoology. The "*Favosites*" mentioned by Whitfield (1883, p. 368) and compared with *F. favosa* may be the specimen figured here.

Geologic location.—Limestone of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimen.—Y.P.M. 20117

Unidentified Crinoid Columnals

(Pl. 3, figs., 4, 5)

Remarks.—Medium-sized crinoid columnals are very abundant in the calcareous quartzite, and large crinoid columnals are abundant in the limestone of the Bernardston Formation. All those in the quartzite are flattened and sheared (Pl. 3, fig. 5), whereas those in the limestone are not nearly so deformed (Pl. 3, fig. 4). Those from the limestone are cited by E. Hitchcock (1833, p. 295, Pl. 14), who called them "encrinites", and by Clarke (*in* Emerson, 1898, p. 260), who compared them to Eaton and Vanuxem's "tricircled encrinite."

Geologic location.—Calcareous quartzite and limestone of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimens.—Y.P.M. 20118, U.S.N.M. 125256

Unidentified Sphaerirhynchiform Brachiopod

(Pl. 3, figs., 10, 11)

Remarks.—Several external impressions of costellate rhynchonellids have the general appearance of *Sphaerirhynchia*, but the material is inadequate for generic identification. The costellate sulcus (Pl. 3, figs. 10, 11), costellate flanks, and nasute outline suggest that this form is a sphaerirhynchiform rhynchonellid. The preservation of relatively delicate ornamentation in the quartzite of the Bernardston Formation is very significant in identification of the abundant associated spiriferoids and will be stressed later.

Similar rhynchonellids are probably the fossils upon which Whitfield (1883, p. 368) based his determination of "two Rhynchonellas . . . one of the species strongly remind one of R. *neglecta* Hall", and upon which Clarke (*in* Emerson, 1898, p. 260) based his determinations of "Well-defined Camarotoechias, like *Rhynchonella sappho* and *R. congregata*, too obscure for specific identification." Boucot believes that none of the rhynchonellids available at the present, or most likely in the past, warrants generic, much less specific, identification.

Geologic location.—Calcareous quartzite of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimens.—Y.P.M. 20119, U.S.N.M. 125265

Unidentified Rostrospiroid Brachiopod

(Pl. 3, figs. 6, 7)

Remarks.—Several internal impressions of pedicle valves appear to belong to an unidentified rostrospiroid. They have short dental plates, a very nasute outline, and the shells were apparently free of any coarse ornamentation. These shells could be assigned to any one of several genera, but the material is too poor for generic assignment.

Similar specimens were probably identified by Whitfield (1883, p. 368) as "casts of a *Terebratuloid* shell, resembling very strongly those of the ventral valve of *Cryptonella eudora*." Although these specimens might be terebratuloids, their association with probable eospiriferids of the *E. radiatus* type suggests otherwise.

Geologic location.—Calcareous quartzite of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimens.—Y.P.M. 20115, 20114

Genus *Eospirifer* Schuchert 1913

Eospirifer cf. *E radiatus* (Sowerby, 1825)

(Pl. 2, figs. 1–6)

Exterior.—Outline subcircular, coarse ornamentation not evident in any of the specimens studied, fine ornamentation probably destroyed during metamorphism.

Pedicle interior.—Massive dental plates on either side of a well-defined muscle field, posterior pedicle callist well defined, no impress of any coarse external ornamentation.

Brachial interior.—Long, slender parallel crural plates support the discrete hinge plates. No evidence of the impress of coarse external ornamentation.

Remarks.—This material is that referred by Whitfield (1883, p. 368) to "a *Spirifera* (quite abundant), resembling in most of its features *Sp. disjuncta* as it occurs in the Chemung Group of New York, but not showing any plications on any

of the internal casts, though impressions of a shell having plications of about the same size and form have been detected in several instances among Professor Clarke's specimens." Clarke (*in* Emerson, 1898, p. 259) suggested that this material might be similar to "*S. granulosus*" or "*S. disjunctus*." *Spinocyrtia granulosa* (Conrad, 1839) can be ruled out as a possibility for the spiriferoid from the Bernardston because *Spinocyrtia* has coarse, radiating costae unlike the smooth ornamentation of the material from the Bernardston. *Cyrtospirifer* Nalivkin, 1918, is ruled out on the same grounds (Pl. 2, fig. 7), because of its coarse, radiating ornamentation that even impresses the interior. The "shell having plications" that Whitfield (1883) refers to was an external impression not seen in association with the characteristic interiors and therefore has no bearing on the identification, whereas the spiriferoid illustrated (Pl. 2, figs. 1, 2) is smooth and consists of an associated interior and exterior. Odd fragments of costellate brachiopod external ornamentation are found in the calcareous quartzite bed (Pl. 3, fig. 8) but they cannot be assigned safely to any of the brachiopod interiors because they do not occur in mutual association.

In Boucot's opinion the spiriferoid from the Bernardston is an eospiriferid of the *E. radiatus* type. Figure 8 of Plate 2 shows a pedicle interior of *Eospirifer* cf. *E. radiatus* from the Silurian Chesuncook Limestone (Willard, 1945) of northern Maine. Note the massive dental plates that after flattening of the shell might well resemble specimens from the Bernardston Formation, and also note the lack of coarse ornamentation. Figures 9 and 10 of Plate 3 show *E. gibbosa* from the Silurian Chicago Formation of Taylor (1930); Figure 9 shows the long, slender, parallel crural plates similar to those shown in material from Bernardston (Pl. 2, fig. 5, 6).

Eospiriferids are known from the Lower Devovonian of eastern North America ("*E.*" *macropleura*), but they are very coarsely plicated and therefore unlike the spiriferoid from Bernardston. Other spiriferoids from the Lower Devonian of eastern North America have costate flanks unlike the smooth flanks of the spiriferoid from Bernardston.

The "*Pentamerus*" noted by Dana (1877, p. 381) is later (*in* Whitfield, 1883, p. 369) assigned by him to the "*Spirifers*" described by Whitfield (1883).

Geologic location.—Calcareous quartzite of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimens.—U.S.N.M. 125261, 125258, 125259; Y.P.M. 20112, 20111

Genus *Leptaena* Dalman 1828

Leptaena? cf. *L*? "*rhomboidalis*" Wilckens

(Pl. 3, fig. 12)

Remarks.—A single impression of a brachial valve exterior has the corrugated ornamentation and geniculation of a strophomenoid like *L.* "*rhomboidalis*", but too little material is available for definite identification. Whitfield (1883) probably had similar specimens upon which to base his "*Strophomena rhomboidalis*."

Geologic location.—Calcareous quartzite of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimen.—U.S.N.M. 125260

Genus *Cyrtina* Davidson 1858

Cyrtina? sp.

(Pl. 3, fig. 9)

Remarks.—A single internal impression of a pedicle valve has the pyramidal form, sulcus, and costellate ornamentation usually associated with *Cyrtina*, but the preservation of the material is too poor for definite assignment. This is probably the material upon which Whitfield (1883) based his determination of a "*Cyrtina* closely resembling *C. Hamiltonensis*." The genus is well known in the Silurian and the Devonian.

Geologic location.—Calcareous quartzite of the Bernardston Formation

Occurrence.—Williams Farm, Bernardston, Mass.

Figured specimen.—Y.P.M. 20113

Pelecypods

Remarks.—Clarke (1909) mentions specimens of *Actinopteria*, *Palaeoneilo*, and *Microdon*, but Boucot found no material that might be construed as the remnants of pelecypods. In this metamorphosed terrane it is very unlikely that thin-shelled pelecypods would survive in a recognizable condition, as even the relatively massive brachiopods are in poor condition. Whitfield (1883) never saw any material that resembled pelecypods, so it seems probable that Clarke misidentified brachiopod fragments as those of pelecypods. Boucot has restudied the material available to both Whitfield and Clarke.

Age of the Calcareous Quartzite of the Bernardston Formation

Evidence from the restudy of the Amherst College "*Spirifer*" referred to by both Balk (1941) and Cooper *et al.* (1942, p. 1747) shows conclusively that it did not come from Massachusetts but is probably from the Nictaux area of Nova Scotia. Therefore for dating the rocks at Bernardston we must rely on the fossils obtained from the calcareous quartzite and the limestone at Bernardston. Most of these fossils—crinoid columnals, tetracorals, rhynchonellids, a possible rostrospiroid, and a questionable *Leptaena*—prove only that they are post-Early Ordovician. However, the fauna also contains an eospiriferid similar to *E.*

radiatus and a possible cyrtinid. These indicate an age younger than middle Early Silurian (Middle Llandovery or Thorold). The eospiriferid allows a more precise dating of late Early Silurian to early Late Silurian (Clinton to Lockport-Guelph; Upper Llandovery to Lower Ludlow). *E. radiatus*, as far as it is known in the northern Appalachians and adjacent parts of New York and Pennsylvania, is restricted to the Silurian. Thus there is no longer any support for a Devonian age for this part of the Bernardston Formation, but there is good evidence that it is of Silurian age.

Age of the Limestone of the Bernardston Formation

The only fossils recovered from the limestone, favositids and crinoid columnals, indicate a post-Early Ordovician age.

Fossils from the Upper Part of the Clough Formation at Skitchewaug and Croydon Mountains

In 1950, Thompson and Boucot found fossils on the east side of Skitchewaug Mountain, near Springfield, Vermont (Thompson, 1952, p. 18, 19, 64). The fossils are very poorly preserved casts and molds in a highly deformed quartzite of the Clough Formation. Crinoid columnals are abundant, as are linear streaks of lime that suggest, but cannot be demonstrated to have been, brachiopods. The tetracorals and crinoidal debris prove that the quartzites here assigned to the Clough Formation (Thompson, 1954, p. 37–41) are of post-Early Ordovician age. The age determinations based on tetracorals made by Stumm, Easton, and Duncan indicate that a post-Ordovician and pre-Carboniferous age is preferred. It is obvious from their reports that the quality of the tetracorals is too poor for a more definitive assignment.

In northern Maine unmetamorphosed strata of Silurian and Devonian age are lithologically similar to the metamorphosed rocks of the same age in western New England. Study of the tetracoral faunas of the rocks in Maine may provide information that would enable the paleontologist to establish more positively the age of the upper part of the Clough Formation at Skitchewaug Mountain.

Stumm (1951, written communication): "Skitchewaug Mountain. Unidentifiable zaphrentoid corals; best guess, Devonian."

Stumm (1952, written communication): "The specimen [Pl. 4, fig. 7] is probably a representative of a species of *Heterophrentis*, which if true, would indicate a Devonian age for the sediments in question. However, identifying horn corals from molds of the calyx is a hazardous undertaking and I would not like to be considered dogmatic with regard to the above assignment."

Easton (1953, written communication): "Although the corals you showed me from New England are too deformed to permit even generic identification, one specimen (Pl. 4, fig. 7) retains enough features to allow a rough age determination to be made. This specimen is large, has a large cardinal fossula, numerous free major-septa and short-minor septa, a rather thick theca, and no columella. Traces of carinae visible on the sides of septa slope axially and apically at a low angle to the horizontal. To my knowledge carinate corals are all post-Ordovician. Almost all of the corals with carinate septa known to me in post-Devonian strata have prominent columellas and those that don't are quite small organisms whose septa tend to be fused into a wall bounding the fossula."

"The specimen [Pl. 4, fig. 7] is one of the Zaphrenthidae."

"It seems almost certain that your coral is either Silurian or Devonian in age. Inasmuch as the great flare of true zaphrenthids occurred in Devonian time, it follows that your coral is quite probably of Devonian age."

Easton (1955, written communication): "This [Pl. 4, figs. 1, 2] is probably a zaphrenthid coral. The seemingly streptelasmoid twisting of the septa may be due to secondary deformation."

Helen Duncan of the U. S. Geological Survey examined the tetracorals and submitted the following report (memorandum, 1955):

> "The corals exhibit features that almost certainly were not developed in Ordovician time, and none of the specimens look to me like characteristic Carboniferous forms. Knowledge of Silurian horn corals in North America is highly inadequate, but the assemblage does not seem to contain tryplasmids and cystiphyllids, which are commonly abundant in known Silurian faunas. The suggestion that these molds are probably zaphrentoids seems to be about as close a relationship as can be inferred. If this interpretation is correct, Silurian or Devonian age would be indicated, the possibility of Devonian age being slightly favored."

A possible fragment of a dalmanitid pygidium is figured (Pl. 4, fig. 8), but the identity of this specimen is in doubt. Thompson (1954) mentions brachiopods and cephalopods, but the specimens are too deformed and distorted for positive identification.

Although brief and tantalizing, the information gleaned from the upper part of the Clough Formation at Skitchewaug Mountain is the only dating for that formation based on

fossil evidence, except for the information obtained in the unit's type area (Billings and Cleaves, 1934, p. 413–415). It is to be hoped that further collecting and study may help to date this faunule more definitively.

Geologic location.—Upper part of Clough Formation

Occurrence.—Road cut 3.2 miles north of west end of Cheshire Toll Bridge on Route 5 Claremont quadrangle, Vermont

Figured specimens.—U.S.N.M. 203729, 125250, 125262, 125286, 125243, 125249

Tetracorals (Pl. 4, fig. 9) found later by Thompson (in 1955 with F. R. Boyd) on Croydon Mountain near Claremont, New Hampshire, are believed by Thompson to be in the same stratigraphic position as the Skitchewaug Mountain fossils. The corals at Croydon Mountain are not so well preserved as those at Skitchewaug Mountain but resemble them in manner of occurrence.

Conclusions

The upper part of the Clough Formation at Skitchewaug Mountain, near Springfield, Vermont, and on Croydon Mountain, near Claremont, New Hampshire, contains fossils of Silurian or Devonian age. The lower part of the Bernardston Formation in Massachusetts contains fossils of probable Silurian age. On a lithologic basis it is possible that the lower part of the Bernardston Formation, including the conglomerates, quartzites, and calcareous quartzite, is the lateral equivalent of the Clough Formation and that the limestone associated with the Bernardston Formation is a lateral equivalent of the Fitch Formation. If the Clough Formation at Skitchewaug Mountain were of Devonian age[1] the stratigraphy of the Connecticut Valley region would be complicated by the fact that the Clough and overlying Fitch Formations elsewhere are of Silurian age.

[1] Since the completion of this paper Thompson (1956) has found several additional fossil localities in the upper part of the Clough Formation on Croydon Mountain. The fossils, mainly brachiopods, are being studied by Boucot and are of late Early Silurian age (Clinton, *i.e.*, Upper Llandovery). The Clough Formation at Skitchewaug Mountain, then, which is lithologically similar, is possibly also of Silurian age. Nevertheless, the tetracorals found at Skitchewaug still form the factual basis for the dating of the Clough Formation at that locality.

References Cited

Bain, G. W., 1936, Mechanics of metasomatism: Econ. Geology, v. 31, p. 505–526

Balk, Robert, 1941, Devonian Bernardston Formation of Massachusetts restudied (Abstract): Geol. Soc. America Bull., v. 52, p. 2009–2010

—— 1956, Bedrock geology of the Massachusetts portion of the Bernardston quadrangle, Massachusetts-Vermont; U. S. Geol. Survey Geol. Quad. Map GQ 90

Billings, M. P., 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. America Bull., v. 48, p. 463–566

Billings, M. P., and Cleaves, A. B., 1934, Paleontology of the Littleton area, New Hampshire: Am. Jour. Sci., v. 28, p. 412–438

Clarke, J. M., 1909, Early Devonian history of New York and eastern North America: N. Y. State Museum Mem. 9, pt. 2, 250 p.

Cooper, G. A., and others, 1942, Correlation of the Devonian sedimentary formations of North America: Geol. Soc. America Bull., v. 53, p. 1729–1793

Dana, J. D., 1873, On rocks of the Helderberg era in the valley of the Connecticut: Am. Jour. Sci., v. 3, p. 339–352

—— 1877, Note on the Helderberg Formation of Bernardston, Mass., and Vernon, Vermont: Am. Jour. Sci., v. 3, p. 379–387

Emerson, B. K., 1898, Geology of Old Hampshire Co., Mass., comprising Franklin, Hampshire, and Hampden counties: U. S. Geol. Survey Mon. 29, 790 p.

Hitchcock, C. H., 1877, The geology of New Hampshire, part II, Stratigraphical geology: Concord, 684 p.

Hitchcock, Edward, 1833, Report on the geology, mineralogy, botany and zoology of Massachusetts: Amherst, 702 p.

—— 1852, On the geological age of the clay slate of the Connecticut Valley in Massachusetts and Vermont: Am. Assoc. Adv. Sci. Proc., v. 6, p. 299–300

Hitchcock, Edward, and Hitchcock, C. H., 1861, Report on the geology of Vermont: Claremont, N. H., 982 p.

Minguzzi, Carlo, 1950, I costituenti minori dei minerali di ferro elbani: Atti Soc. Toscana Sci. nat. (Pisa) Mem. 57, ser. a., p. 119–144

Palache, Charles, Berman, Harry, Frondel, Clifford, 1945, Dana's manual of mineralogy, v. 1, 7th ed.: New York, John Wiley and Sons, Inc., 834 p.

Scupin, H., 1900, Die Spiriferen Deutschlands: Palaeont. Abhandlungen, N. F., Bd. LV, Heft 3, p. 207–344

Taylor, D. O., 1930, A new shale and related structure in the Chicago area: Ill. Acad. Sci. Trans., v. 22, p. 473–477

Thompson, J. B., Jr., 1952, Southern Vermont: Guidebook for field trips in New England: Geol. Soc. America, p. 14–23, 38–41, 56–65

—— 1954, Structural geology of the Skitchewaug Mountain area, Claremont quadrangle, N.H.-Vt.: New England Intercollegiate Geol. Conf. Guidebook, 46th Conf., Dartmouth College, Hanover, N. H.

Thompson, J. B., 1956, Skitchewaug nappe, a major recumbent fold in the area near Claremont, N. H. (Abstract): Geol. Soc. America Bull., v. 67, p. 1826–1827

Whitfield, R. P., 1883, Observations on the fossils of the metamorphic rocks of Bernardston, Mass.: Am. Jour. Sci., v. 3, p. 368–369

Willard, Bradford, 1945, Silurian fossils from Ripogenus Dam, Maine: Jour. Paleontology, v. 19, p. 64–68

U. S. Geological Survey and Dept. Geology and Geophsics, Mass. Inst. Technology, Cambridge, Mass.; Inst. Geophysics, Univ. of California, Los Angeles, Calif.; U. S. Geological Survey, Washington, D. C.; Dept. Mineralogy and Petrography, Harvard University, Cambridge, Mass.

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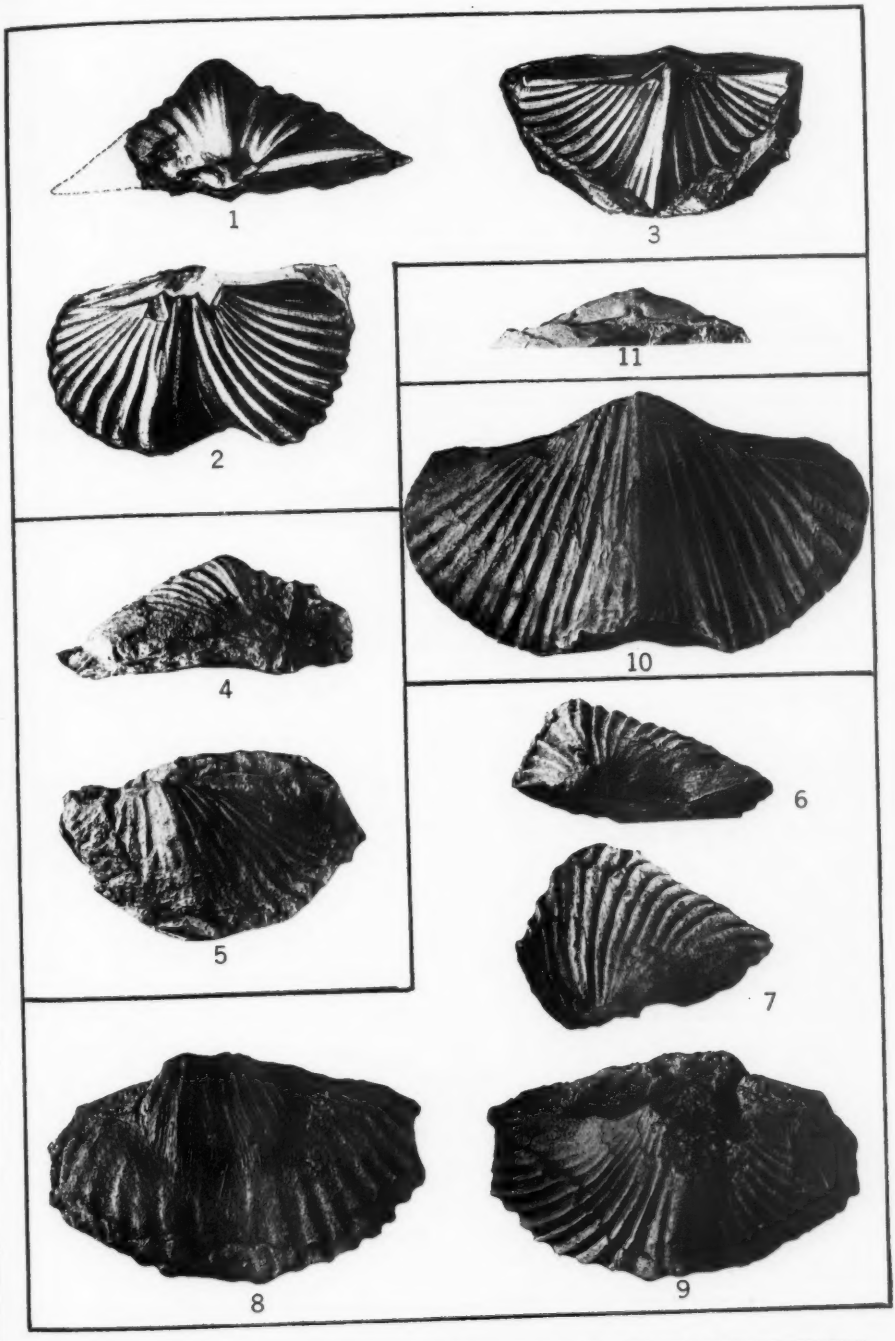
Plate 1.—Lower Devonian Brachiopods from Germany, Nova Scotia, and Kentucky

Figures 1–3.—*"Spirifer" bischofi* Roemer 1858 (1) Posterior view of internal impression of brachial valve (×1). Note absence of crural plates. Upper Coblenzian; Waxweiler, Rhineland. From Scupin (1900, pl. VII, fig. 4b). (2) Internal impression of pedicle valve (×1). Note nonarrectiform nature of muscle field as contrasted with figure 8, and plications in sulcus. Upper Coblenzian; Waxweiler, Rhineland. From Scupin (1900, pl. VII, fig. 2). (3) Internal impression of brachial valve (×1). Note plicated fold and angular plications. Siegen graywacke; Menzenberg, Rhineland. From Scupin (1900, pl. VII, fig. 1).

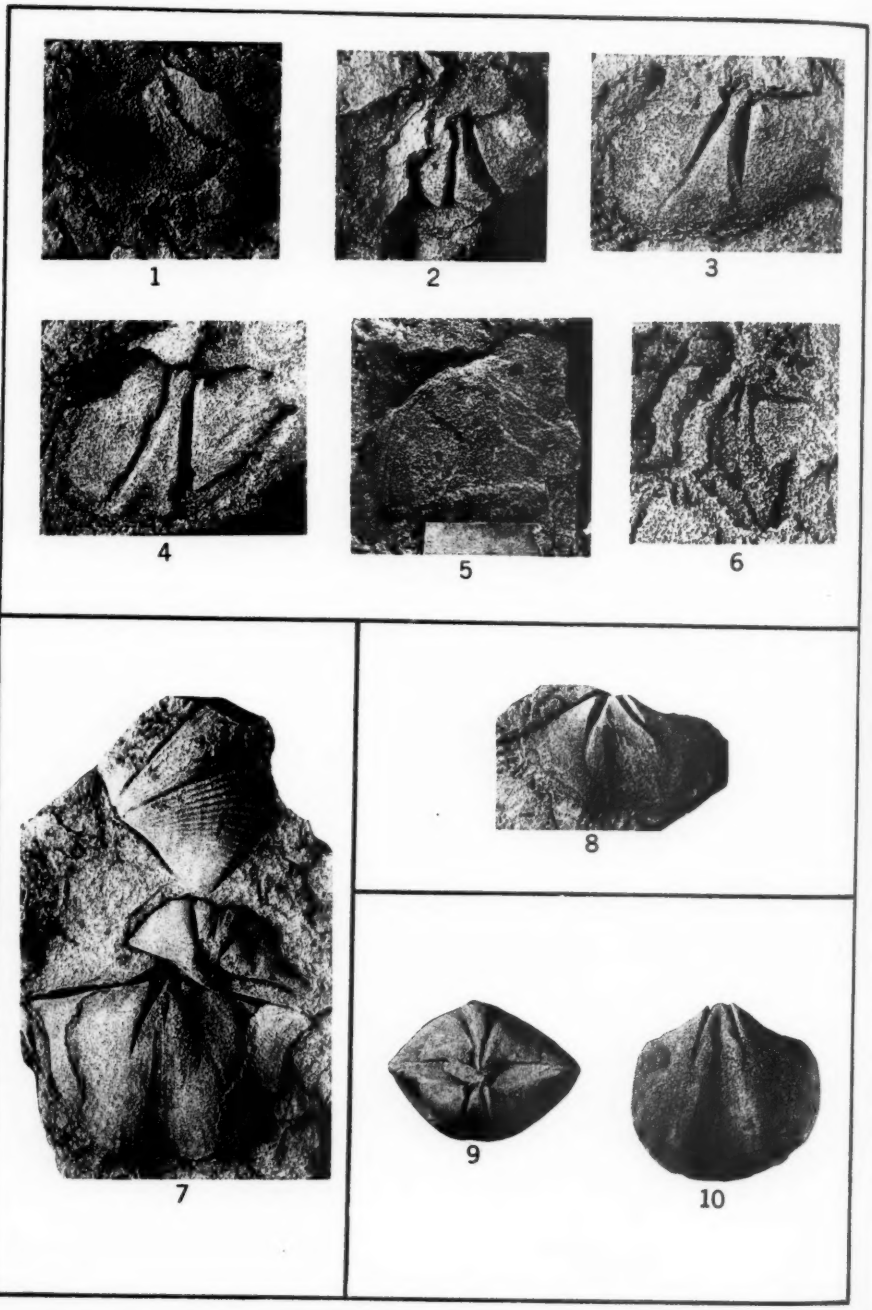
Figures 4–9.—*"Spirifer"* n. gen. et n. sp. (4) Plaster cast, posterior view of internal impression of brachial valve (×1). Note presence of short crural plates. Horizon and locality in question. Probably high Lower Devonian of Nictaux, Nova Scotia. Original in collections of Amherst College. Plaster cast U.S.N.M. No. 125254. (5) Plaster cast of internal impression of brachial valve (×1). Note costate fold and angular plications. Horizon and locality in question. Probably high Lower Devonian of Nictaux, Nova Scotia. Original in collections of Amherst College. Plaster cast U.S.N.M. No. 125254. (6) Posterior view of internal impression of brachial valve (×1). Note presence of short crural plates. High Lower Devonian; Nictaux Falls, Annapolis Co., Nova Scotia. U.S.N.M. No. 125266A. (7) Internal impression of brachial valve (×1). Note costate fold and angular plications. High Lower Devonian; Nictaux Falls, Annapolis Co., Nova Scotia, U.S.N.M. No. 125266A. (8) Internal impression, pedicle view (×1). Note costate sulcus and angular plications. High Lower Devonian; Nictaux Falls, Annapolis Co., Nova Scotia. U.S.N.M. No. 125266B. (9) Internal impression, brachial view (×1). Note costate fold and angular plications. High Lower Devonian; Nictaux Falls, Annapolis Co., Nova Scotia. U.S.N.M. No. 125266B.

Figure 10.—*Fimbrispirifer divaricatus* (Hall, 1857) (10) Brachial view of exterior (×1). Note the costate fold and the angular plications. Limestone of Onondaga age; Lebanon, Kentucky. U.S.N.M. No. 51190A.

Figure 11.—*Fimbrispirifer venustus* (Hall, 1860) (11) Posterior view of internal impression of brachial valve (×1). Note presence of short crural plates. Beechwood Limestone member of Sellersburg Limestone; Lebanon, Kentucky. U.S.N.M. No. 39489A.



LOWER DEVONIAN BRACHIOPODS FROM GERMANY, NOVA SCOTIA, AND KENTUCKY



BRACHIOPODS FROM MASSACHUSETTS, MAINE, PENNSYLVANIA, AND ILLINOIS

PLATE 2.—BRACHIOPODS FROM MASSACHUSETTS, MAINE, PENNSYLVANIA, AND ILLINOIS

FIGURES 1–4.—*Eospirifer* cf. *E. radiatus* (Sowerby, 1825) (1) Impression of pedicle valve exterior (same specimen as Fig. 2) (×1). Note absence of costellae or plications of any sort. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125261. (2) Internal impression of pedicle valve (×1). Note the large, massive dental plates. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125261. (3) Internal impression of pedicle valve (×1). Note the large, massive dental plates. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125258. (4) Internal impression of pedicle valve (×1). Note the large massive dental plates. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125259.

FIGURES 5, 6.—*Eospirifer* cf. *E. radiatus* (Sowerby, 1825). (5) Internal impression of brachial valve (×1). Note the long, narrow crural plates, and the absence of any ornamentation on the internal impression. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20112. (6) Internal impression of brachial valve (×1). Note the long, narrow crural plates. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20111.

FIGURE 7.—*Cyrtospirifer disjunctus* (Sowerby, 1840). Internal impressions of brachial and pedicle valves (×1). Note the impression of the external costae on the internal impressions. Conewango Formation; Tidioute, Pennsylvania. U.S.N.M. No. 70790.

FIGURE 8.—*Eospirifer* cf. *E. radiatus* (Sowerby, 1825). Internal impression of pedicle valve (×1). Note the large, massive dental plates. Chesuncook Limestone of Willard (1945); U.S.G.S. Silurian and Devonian locality 3249, islet about 3 miles southeast from mouth of Quaker Brook, Chesuncook Lake, Piscataquis Co., Maine (field no. Z-270). U.S.N.M. No. 125257.

FIGURES 9, 10.—*Eospirifer gibbosa* (Hall, 1861) (9) Posterior view of internal impression (×1). Note the long, narrow crural plates, as well as the prominent dental plates. Chicago Formation of Taylor (1930); Chicago, Illinois. U.S.N.M. No. 125255B. (10) Internal impression of pedicle valve (×1). Note the large, massive dental plates. Chicago Formation of Taylor (1930); Chicago, Illinois. U.S.N.M. No. 125255A.

PLATE 3.—FOSSILS FROM MASSACHUSETTS

FIGURE 1.—Favositid coral (×1). Note tabulae. Limestone of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20117.

FIGURES 2, 3.—Unidentified tetracorals (2) Internal impression of tetracoral (×2). Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125264. (3) Internal impression of tetracoral (×2). Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125263.

FIGURES 4, 5.—Crinoid columnal (4) Crinoid columnal (×1). Limestone of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20118. (5) Impression of pelmatozoan columnal. Note how flattening and shear have affected this originally cylindrical object. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125256.

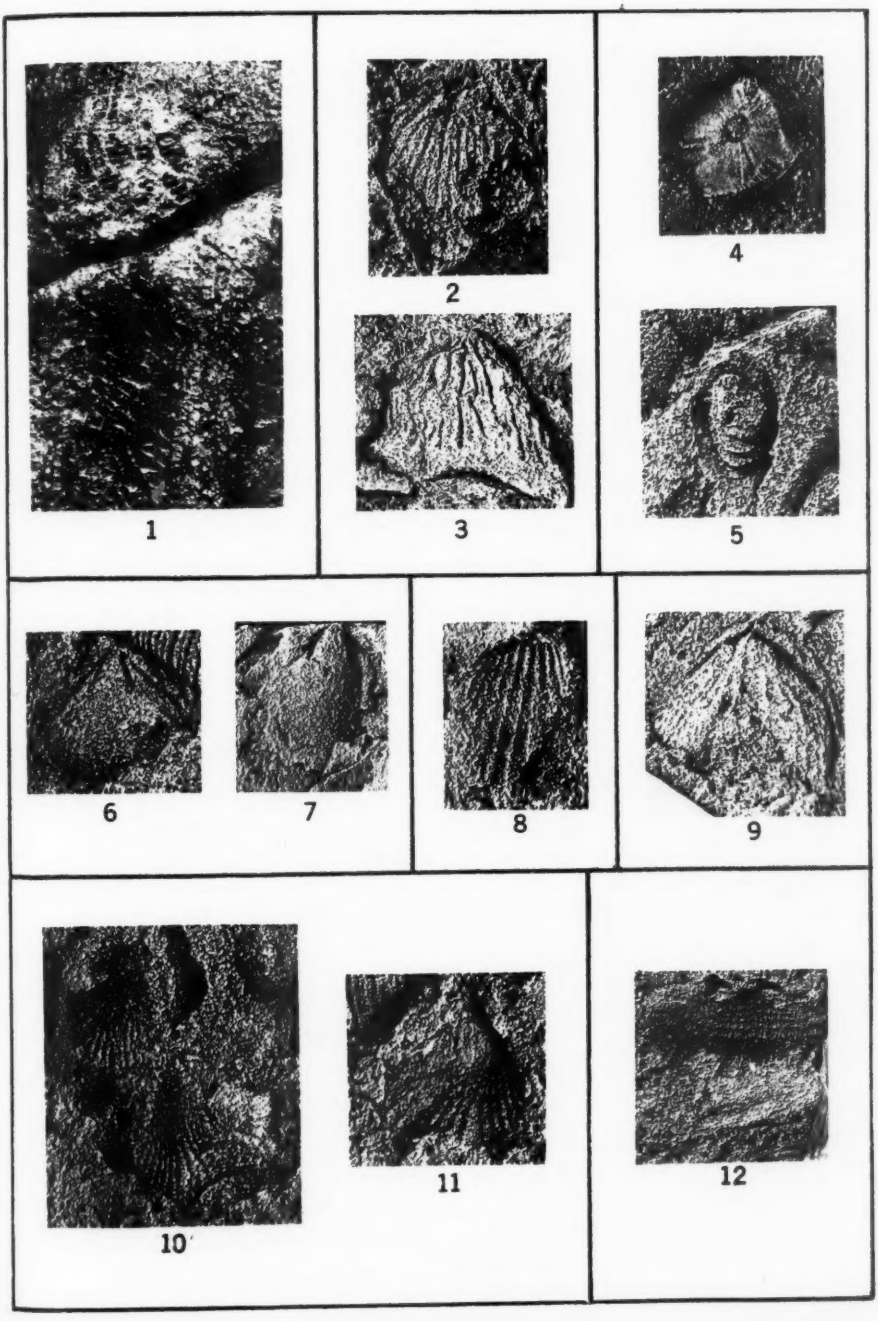
FIGURES 6, 7.—Unidentified rostrospiroid brachiopod (6) Internal impression of pedicle valve (×2). Note the short dental plates and the lack of any ornamentation on the internal impression. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20115. (7) Internal impression of pedicle valve (×1). Note the short dental plates and the lack of any ornamentation on the internal impression. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20114.

FIGURE 8.—Costellate fragment of brachiopod. Impression of external costellae of a brachiopod (×2). Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20116.

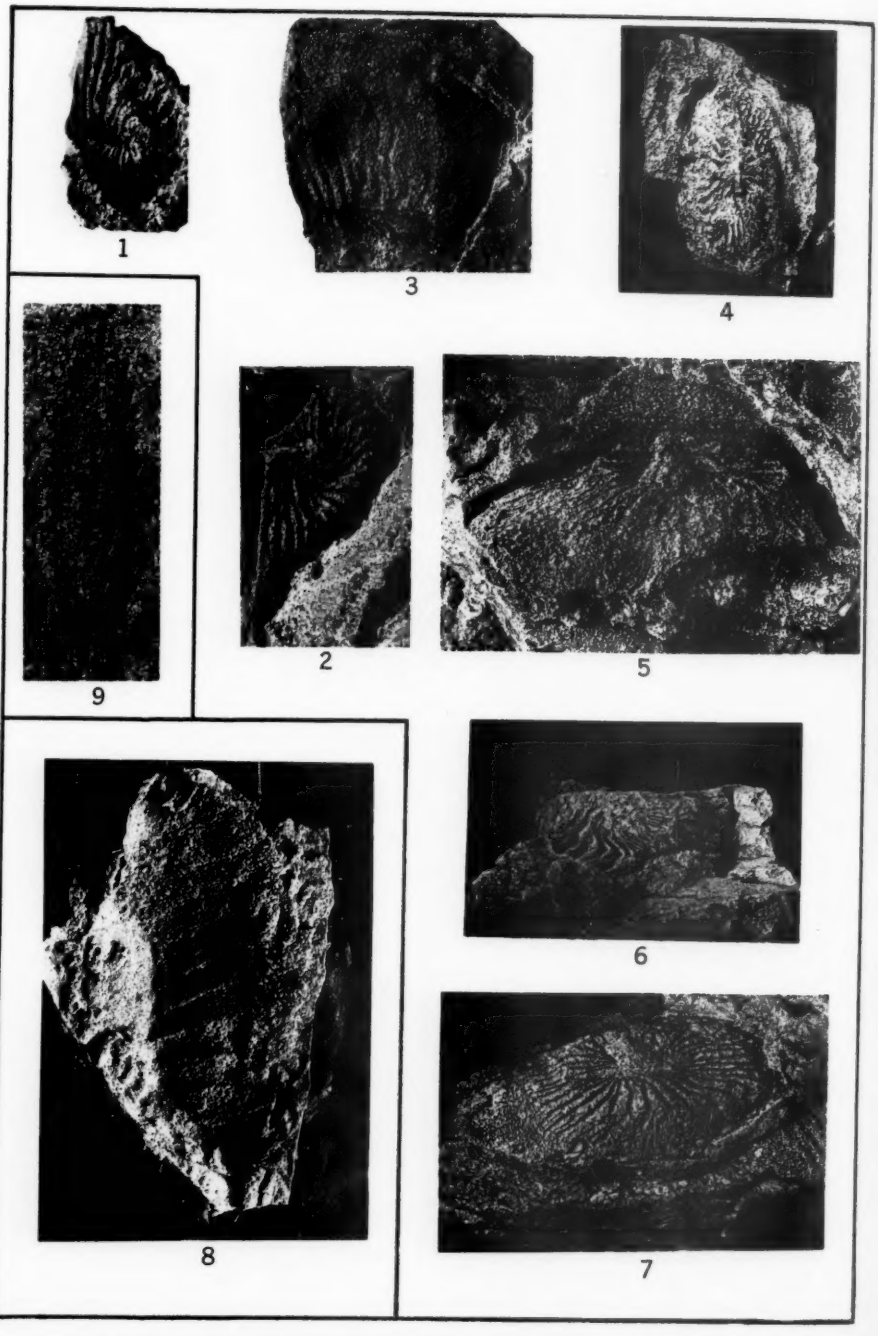
FIGURE 9.—Cyrtiniform brachiopod. Impression of pedicle interior (×2). Note the subpyramidal form and the apparently costellate flanks. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20113.

FIGURES 10, 11.—Sphaerirhynchiform brachiopod (10) Impression of exterior of pedicle valve (×2). Note the costellate sulcus and flanks, and the nasute beak. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. Y.P.M. No. 20119. (11) Impressions of exterior (×2). Note the costellae on both impressions. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125265.

FIGURE 12.—*Leptaena*? cf. *L.*? *"rhomboidalis"* Wilckens. External impression (×1) of brachial valve. Calcareous quartzite bed of Bernardston Formation; Williams Farm, Bernardston, Mass. U.S.N.M. No. 125260.



FOSSILS FROM MASSACHUSETTS



FOSSILS FROM EASTERN VERMONT AND WESTERN NEW HAMPSHIRE

PLATE 4.—FOSSILS FROM EASTERN VERMONT AND WESTERN NEW HAMPSHIRE

FIGURES 1–7.—Unidentified zaphrentid tetracorals (1) Rubber mold of calyx (×2). Note the fossula. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. Acc. 203729. (2) Impression of calyx (×2). Note impression of fossula. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. Acc. 203729. (3) Impression of calyx (×1). Note the flattening caused by deformation. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125250. (4) Impression of calyx (×1). Note the elliptical outline due to shear. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125262. (5) Impression of calyx (×1). Note the extreme flattening caused by deformation. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125286. (6) Impression of calyx (×1). Note the sigmoid curvature of the septae due to rotation during deformation. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125243. (7) Impression of calyx (×1). Note the elliptical outline due to shear. Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125249.

FIGURE 8.—Possible dalmanitid trilobite pygidium. Rubber mold of possible dalmanitid trilobite pygidium (×1). Upper part of Clough Formation; road cut 3.2 miles north of west end Cheshire Toll Bridge on U.S. Route 5, Springfield, Vermont. U.S.N.M. 125251.

FIGURE 9.—Unidentified tetracoral. Impression of calyx (×2). Note the elongate elliptical outline due to deformation. Quartzite of the Clough Formation near contact with the Fitch Formation in extreme south part of Plainfield Township at bend in trail a little more than half a mile W-N-W of notch between Grantham Mountain and Croydon Peak, New Hampshire. U.S.N.M. 127397.

PLATE 5.—THIN SECTIONS
×75

No. 2. Coarse garnet-mica gneiss. Bernardston, Mass.

The subrounded garnets are in a matrix of olive-brown biotite and calcite (or dolomite) with limonitic veinlet. Some of the garnet contains green chlorite (after biotite?) and is intergrown with carbonate, hematite, and possibly pyrrhotite.

With the matrix biotite is green chlorite of the same type as in Pl. 6, no. 1. The biotite is not detrital, but clearly of metamorphic origin, like the garnet.

No magnetite was seen in the section; the rock is not attracted by a magnet, and no magnetic fraction could be obtained from the crushed rock.

No. 4. Coarsely granular magnetite rock. Bernardston, Mass.

Euhedral magnetite (black) with large garnets (just right of center, etc.). With the magnetite at the left of this garnet is much olive-brown biotite. The small clear areas are quartz.

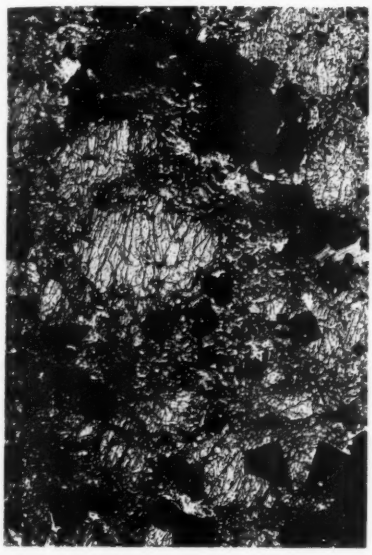
The magnetite shows cleavage, with cubic inclusions of pyrite (Pl. 6, no. 4). The garnet also shows marked shattering, with chlorite, etc. along the cracks. Biotite and chlorite similar to those in the other Bernardston specimens are present; also apatite, calcite, and quartz.

No. 6. Highly fossiliferous magnetite rock. Nictaux, Nova Scotia.

In thin section, very much like No. 7, with additional carbonate. The fossils consist of carbonate. A relatively large fossil is seen near the middle right edge.

No. 7. Magnetite rock. Amherst College specimens

In thin section, rounded quartz grains, clearly detrital, with much secondary chlorite. Chlorite and magnetite, some of the latter apparently replacing fossils, are abundant. Some of the chlorite appears to replace glauconite. The fine-grained area (lower right margin) is recrystallized shale cut by quartz veinlets.



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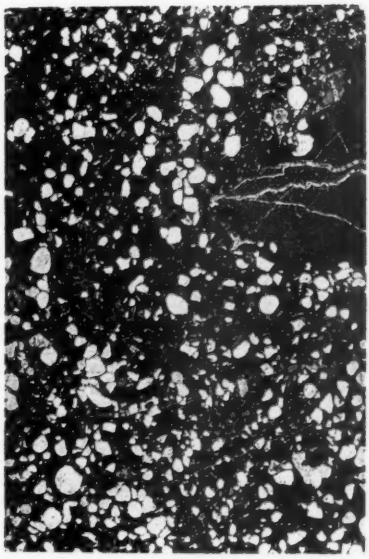

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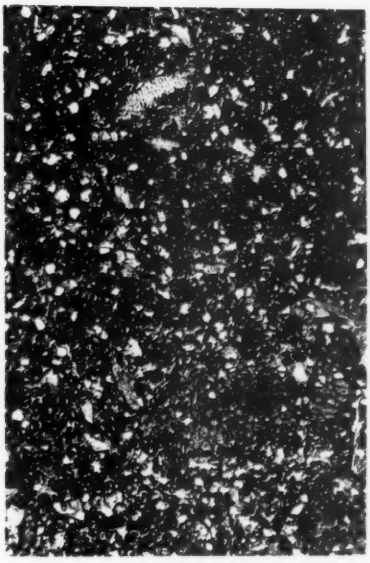
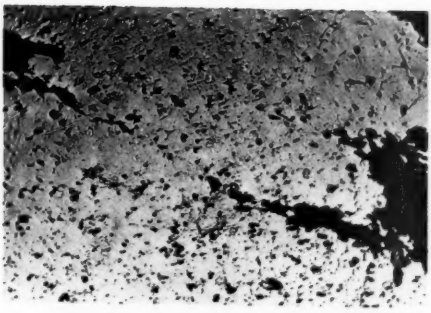


6

1 inch

THIN SECTIONS



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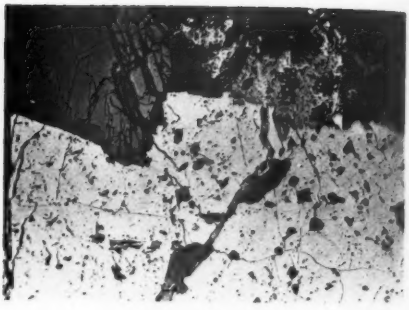
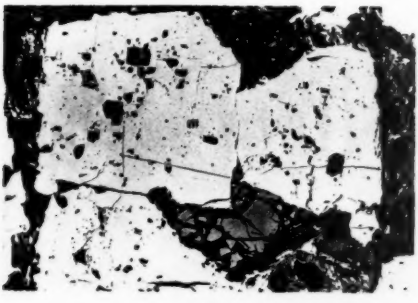
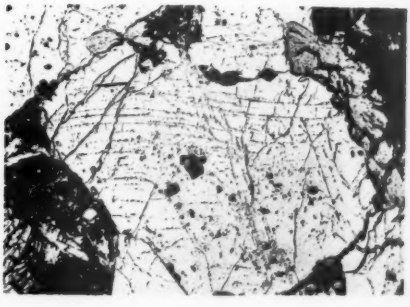


3(a)



3(b)



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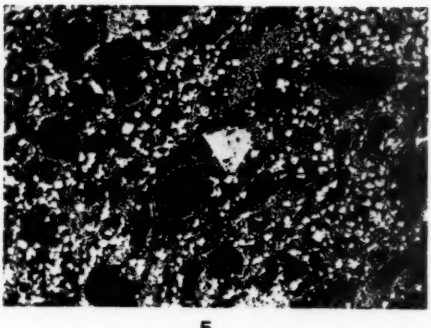

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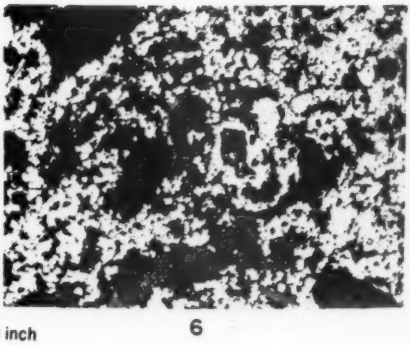

6

1 inch

POLISHED SECTIONS

PLATE 6.—POLISHED SECTIONS
×60

No. 1. Massive magnetite, Bernardston, Mass.

This is massive, fine-grained magnetite and "graphic" hematite intergrowths, with very sparing pyrrhotite (?). In thin section (not illustrated) the gangue shows irregular bands of bright-green chlorite, strongly pleochroic in green and pale tan, with much apatite and quartz. In the quartz is scattered prismatic carbonate (calcite?).

No 3. Magnetite rock, Bernardston, Mass.

(a) In polished section, abundant coarse garnet (upper left) and magnetite (lower half). Limonite (goethite) is abundant, in part replacing magnetite, in part pyrite (upper right center).

In thin section much like no. 2, except for its magnetitic content (absent in 2).

(b) Magnetite with cubic pyrite inclusions, mostly altered to limonite. Garnet, lower center (dark).

No. 4. Magnetite rock, Bernardston, Mass.

Showing deformed cleavage in magnetite.

No. 5. Magnetite rock, Nictaux, Nova Scotia.

Large magnetite crystal in center, with numerous smaller ones. The bright granular material (upper center right, etc.) is hematite. Bright iron-oxide particles also surround fossils.

In thin section (not illustrated) abundant detrital particles, including quartz with chlorite, siderite (?), chert (?), and possible volcanic rock. There is also a fragment of sandstone. There are also fossils, rather poorly preserved. Unlike the other specimens, the rock is only slightly metamorphosed.

No. 6. Magnetite in fossiliferous rock, Nictaux, Nova Scotia.

At the center, surrounded by gangue, are two specks of chalcopyrite (?), the only such seen in these specimens.

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PALEOECOLOGY OF THE CARBONIFEROUS REEF COMPLEX AND SHELF LIMESTONES IN NORTHWEST DERBYSHIRE, ENGLAND

BY E. BRIAN WOLFENDEN

ABSTRACT

True reef limestones discovered in the mid-Visćan (Upper Mississippian) reef complex of northwest Derbyshire are composed for the most part of massive calcareous algae of the *Collenia-Cryptozoon* types. The reefs developed at two horizons and are not continuous either in time or space. They developed at the margin of a shallow shelf and formed an incipient barrier reef which separated the open seas of the English Midlands from a shallow shelf sea. Each reef was elongate parallel to the margin of the shelf and was up to 1200 feet long. The reefs averaged 30 feet wide and accumulated in thicknesses up to 100 feet. At the shelf margin in front of the reefs was a steep slope leading to the adjacent basin floor which was as much as 400 feet below the reef surface.

The shelf limestones are calcirudites, calcarenites, and calcilutites. Spergenites, which formed in shallow, warm sea water supersaturated with calcium carbonate, compare closely with sediments on the Bahama Banks. Aphanitic limestones within the reef complex possibly represent recrystallized algal deposits.

Few species lived in the shelf province, but the reef-complex fauna was rich both in species and individuals. The reef—built for the most part of algal deposits—had a distinctive sponge and bryozoan fauna.

Identical reefs occur in the Craven reef complex of Yorkshire, but no such facies occurs in *C* zone reef knolls of Bowland, Lancashire.

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Introduction and Acknowledgments

This paper describes an investigation of the environmental conditions which led to the formation of the reef complex and shelf limestones of mid-Viséan age in northwest Derbyshire, England (Table 1; Fig. 1). It is based on detailed field work, with special attention to the facies distribution of the fossils and to the lithology of the limestones, particularly as seen in thin sections.

A reef is here taken to be a sedimentary rock aggregate, composed for the most part of the *in situ* remains of colonial-type organisms that lived at shallow depths in seas (Wilson, 1950, p. 181). Following Henson (1950) and Newell *et al.* (1953, p. 48), those limestones previously referred to as "reef" are here considered to represent an incipient reef complex, within which limestones made up mainly of algal deposits are believed to represent the reef. The reefs are not continuous either in time or space. They developed at two horizons; the lower reef is of lower B2 age, and the upper one is of upper B2 age. Each reef forms elongate banks as much as 1200 feet long and about 30 feet wide, which parallel the margin of the shelf province. The reefs accumulated in thicknesses up to 100 feet; the height to which they grew above the surrounding shelf floor is not known. Limestones genetically associated with the reef are termed fore reef and back reef, depending on their position; they can be distinguished only when the reefs are developed.

Field relationships and paleontologic evidence indicate that the B2 subzone can be correlated with strata in D1 from just below the *Cyrtina* band to the top of the subzone. The reef complex, of B2 age, has been examined at Earl Sterndale and Castleton. Strata in the shelf limestones from the *Cyrtina* band to the top of the D1 subzone (Fig. 2) have been examined behind the reefs in these two areas and almost continuously from Earl Sterndale to Miller's Dale, a distance of 5 miles perpendicular to the trend of the reef complex (Figs. 3 and 4).

The reef complex at Castleton has been investigated by Shirley and Horsefield (1940) and Parkinson (1943; 1947; 1953) among others. Shirley and Horsefield consider that the B2 limestones rest unconformably on both D1 and D2 shelf limestones; Parkinson, however, considers B2 equivalent to D1 and attributes the difference in fauna and lithology to a facies change.

The author is indebted to Dr. I. M. Simpson and Dr. S. H. Straw for encouragement during the research and preparation of the paper. Dr. F. W. Anderson kindly confirmed the identification of algae in the reef and allowed the author to study a collection of slides of Carboniferous algae. The author is indebted to Dr. W. W. Black, Prof. F. W. Cope, and Dr. D. Parkinson for valuable discussion in the field. The author gratefully acknowledges a grant from the Department of Scientific and Industrial Research.

Stratigraphy

Shelf Province

The shelf province includes those areas behind the reef. The shelf limestones that have been examined are remarkably uniform across the shelf, but lithologic and paleontologic variations have been observed within about a quarter of a mile of the reef.

Cope (1933) recognized in the Miller's Dale area a succession of limestones ranging from the S2 to D2 subzones; the base of these limestones is not exposed. The present investigation has been confined to strata from the *Cyrtina* band to the top of the D1 subzone (Fig. 2). The same succession described by Cope (1939) at Miller's Dale occurs at Castleton (Shirley and Horsefield, 1940) and has now been recognized at Earl Sterndale.

The succession examined begins with the thickly bedded, pale-gray, pure, clastic Chee Tor limestones. Partial recrystallization has in places resulted in the formation of pseudobreccias (Wood, 1941). Chert is absent. Fossils are uncommon, but the *Cyrtina* band 25 feet from the top contains many fossils. The following fossils occur in these limestones:

Caninia sp., *Chaetetes depressus* (Fleming), *Dibunophyllum bourtonense* Garwood and Goodyear,

Lithostrotion irregulare (Phillips), *L. junceum* (Fleming), *Palaeosmilia murchisoni* Edwards and Haime, *Syringopora* sp., *Athyris* sp., *Daviesiella* aff. *comoides* (J. Sowerby), *Linoproductus* cf. *hemisphericus* (J. Sowerby), *Bellerophon* sp., crinoids, foraminifers, ostracods, and algae including *Calcisphaera* sp., *Girvanella* sp., and *Koninckopora inflata* (de Koninck).

Within about a quarter of a mile of the reef the limestones vary both in fauna and lithology.

Limestones with oölitic grains occur at Earl Sterndale and Castleton. At Castleton, such limestones occur in the Winnats Pass and on Treak Cliff in both Chee Tor and Miller's

TABLE 1.—STRATIGRAPHIC CLASSIFICATION OF ENGLISH LOWER CARBONIFEROUS ROCKS WITH EQUIVALENT AMERICAN STAGES

		Coral-brachiopod zones		Goniatite zones		American stages
Dinantian (Mississippian)	Viséan	D	D2		P1	Chester
			D1	B	B2	
		S	S2		B1	Meramec
			S1			
		C	C2			
			C1			Osage
	Tournaisian	Z				Kinderhook
		K				

The basalt Lower Lava Flow on top of these limestones has a maximum thickness of 75 feet. It has a discontinuous outcrop, probably the consequence of normal thinning of the flow rather than of penecontemporaneous erosion.

Resting on this lava flow are the Miller's Dale limestones, which are similar to the Chee Tor limestones. They are thickly bedded, pale-gray, pure, clastic rocks and in many places are fine-grained. Many of these limestones contain abundant fossils belonging to only a few species. About 20–30 feet above the base is the *Lithostrotion m'coyanum* band, which is characterized by numerous coral colonies and has been recognized in many areas. The following fossils occur in these limestones:

Caninia benburbensis Lewis, *Chaetetes septosus* (Fleming), *Dibunophyllum* cf. *bipartitum* (M'Coy), *D. bourtonense*, *Lithostrotion basaltiforme* (Phillips), *L. irregulare*, *L. junceum*, *L. m'coyanum* Edwards and Haime, *Palaeosmilia murchisoni*, *Athyris* sp., *Chonetes papilionacea*, *Davidsonina septosa*, *Gigantoproductus maximus* (M'Coy), *Bellerophon* sp., crinoids, foraminifers, ostracods, and algae including *Calcisphaera* sp., *Girvanella* spp., and *Koninckopora inflata*.

Dale limestones. Outcrops from Cave Dale and south of Windy Knoll have been described by Parkinson (1947), under the term "beach beds". The horizon in both places was about 30 feet above the Lower Lava Flow. Oölitic limestones also occur south of Snels Low and Middle Hill near the *Cyrtina* band. At Earl Sterndale, they are found at Dowel Dale, Upper Edge, Jericho, Aldery Cliff, and south of High Needham, and they are developed at several horizons. At both Earl Sterndale and Castleton they are most common in the *Cyrtina* band.

The *Cyrtina* band has been recognized only 20 yards behind the reef in Dowel Dale, where, together with normal shelf limestones, are aphanitic limestones of back-reef facies. The latter contain bands of fibrous calcite crystals, which W. W. Black (1952, p. 195) has shown are secondary. Together with the fauna normal to the band, the following reef-phase fossils have been found:

Athyris spp., *Dielasma hastata* (J. de C. Sowerby), *Gigantoproductus* sp., *Pugnax pugnus* (Martin), *Schizophoria resupinata* (Martin), *Spirifer bisulcatus* J. de C. Sowerby, *S.* sp., *Straparollus* spp.

In the Winnats Pass, near the reef, the following reef-phase fossils occur in the band:

Dielasma hastata, Pugnax pugnus, Schizophoria resupinata, Spirifer bisulcatus.

Castleton, on Treak Cliff and the Winnats Pass (Fig. 6), and from Perryfoot to near Windy Knoll (Fig. 7).

In the Earl Sterndale area, reef limestones are exposed in ridges which extend up the

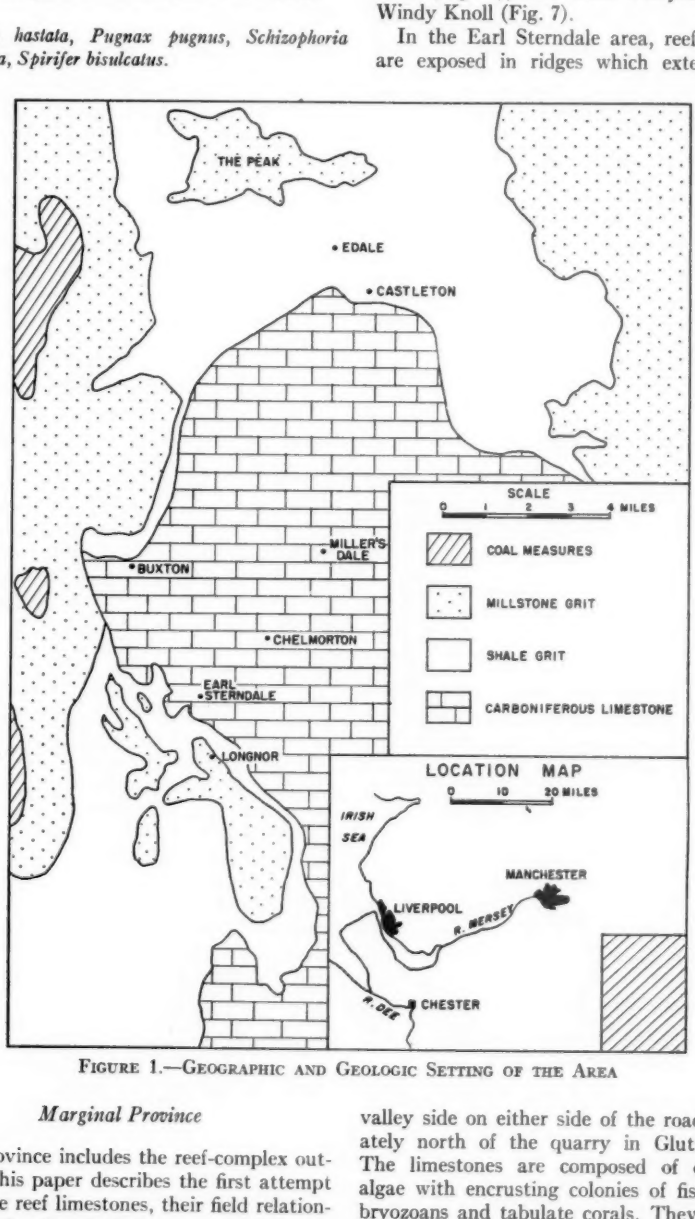


FIGURE 1.—GEOGRAPHIC AND GEOLOGIC SETTING OF THE AREA

Marginal Province

This province includes the reef-complex outcrop. As this paper describes the first attempt to map the reef limestones, their field relationships will be described in some detail. Reef limestones are exposed at Earl Sterndale, from Chrome Hill to Aldery Cliff (Fig. 5); at valley side on either side of the road immediately north of the quarry in Glutton Dale. The limestones are composed of calcareous algae with encrusting colonies of fistuliporoid bryozoans and tabulate corals. They crop out as an almost vertical wall 80 feet high and then die out. The reef averages 30 feet in width. North of the reef are unbedded apha-

nitic limestones of back-reef facies with a rich brachiopod fauna. These limestones are seen immediately above the road level 100 yards north of the reef. Above them are almost-horizontal shelf limestones which grade to the south into reef-complex limestones. South of the reef, in the quarry, are steeply

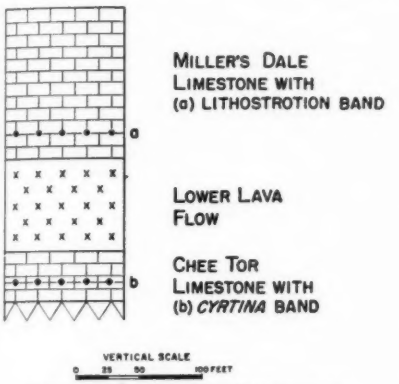


FIGURE 2.—SUCCESSION EXAMINED IN THE SHELF LIMESTONES

dipping fore-reef limestones with a rich fauna of brachiopods and mollusks, together with *Davidsonina septosa*, common *Palaeosmilia murchisoni*, *Dibunophyllum bourtonense*, rare *Palaeosmilia regia* (Phillips), and a few goniatites of lower B2 age; this fauna represents the *Beyrichoceras vesiculiferum* subzone of Hudson and Cotton (1944). These limestones are not typical of the fore-reef facies, in that they contain common shell debris; also, a breccia bed composed of blocks of reef limestone in a friable matrix occurs in front of the reef.

This reef is exposed in the valley west of Glutton Dale; it crops out there as a 30-foot, almost-vertical wall and dies out at an altitude of 1000 feet. To the north are back-reef limestones with a fauna of brachiopods and gastropods, together with lower B2 goniatites. The *Cyrtina* band occurs at an altitude of 1020 feet in shelf limestones 200 yards northwest of the reef. South of the reef are steeply dipping fore-reef limestones. Above the reef, shelf limestones pass laterally into reef-complex limestones. The position at which the facies change is accomplished passes to the south at progressively higher horizons. At these transition horizons, almost-horizontal shelf limestones can be seen directly above the reef. South of the reef, they overlie steeply dipping fore-reef limestones.

A reef at this same horizon is developed in Dowel Dale. As elsewhere it crops out as an almost-vertical wall. The reef is 50 feet high and 30 feet wide. South of this are fossiliferous fore-reef limestones containing common lower B2 goniatites. North of the reef the *Cyrtina* band is exposed, and a specimen of *Davidsonina septosa* has been obtained directly above the reef, in reef-complex limestones. The reef is best developed west of the road that runs through the valley.

The upper reef is found well to the south of the lower reef. This upper reef is exposed on Chrome Hill, Parkhouse Hill, and Aldery Cliff.

On Chrome Hill the upper reef is seen east of the summit on the steep northern slopes, where it crops out as an almost-vertical wall about 30 feet wide. From this outcrop it can be traced west into gullies that run south from the summit of the hill. Exposures in these show the transition from shelf limestones in the north through poorly bedded back-reef limestones, then reef limestones, to steeply dipping fore-reef limestones. These last contain upper B2 goniatites; this is the *Beyrichoceras delicatum* subzone of Hudson and Cotton (1944). In the most western outcrop the reef appears to die out laterally.

On Parkhouse Hill this reef is exposed on the steep northern slopes, both in the cliffs below the summit and in smaller exposures to the east. The reef is 70 feet high; the top of the hill is at an elevation of 1250 feet. As elsewhere the reef crops out as an almost-vertical wall, the width of which is unknown, because the equivalent back-reef limestones have been eroded. South of the reef, steeply dipping, fossiliferous fore-reef limestones contain plentiful goniatites of upper B2 age.

The part of the reef exposed for about 400 yards on Aldery Cliff averages 60 feet high and 30 feet wide. Traced to the east and west it appears to die out laterally. To the north, interbedded back-reef and shelf limestones occur, and to the south are steeply dipping fore-reef limestones of upper B2 age.

The development of the reef complex in the Earl Sterndale area is seen in the idealized section in Figure 5.

In the Castleton area, on Treak Cliff and in the Winnats Pass, a comparable development of reefs can be demonstrated (Fig. 6).

As at Earl Sterndale, a lower reef developed which grew almost vertically upward with a width of about 30 feet. It is well exposed on

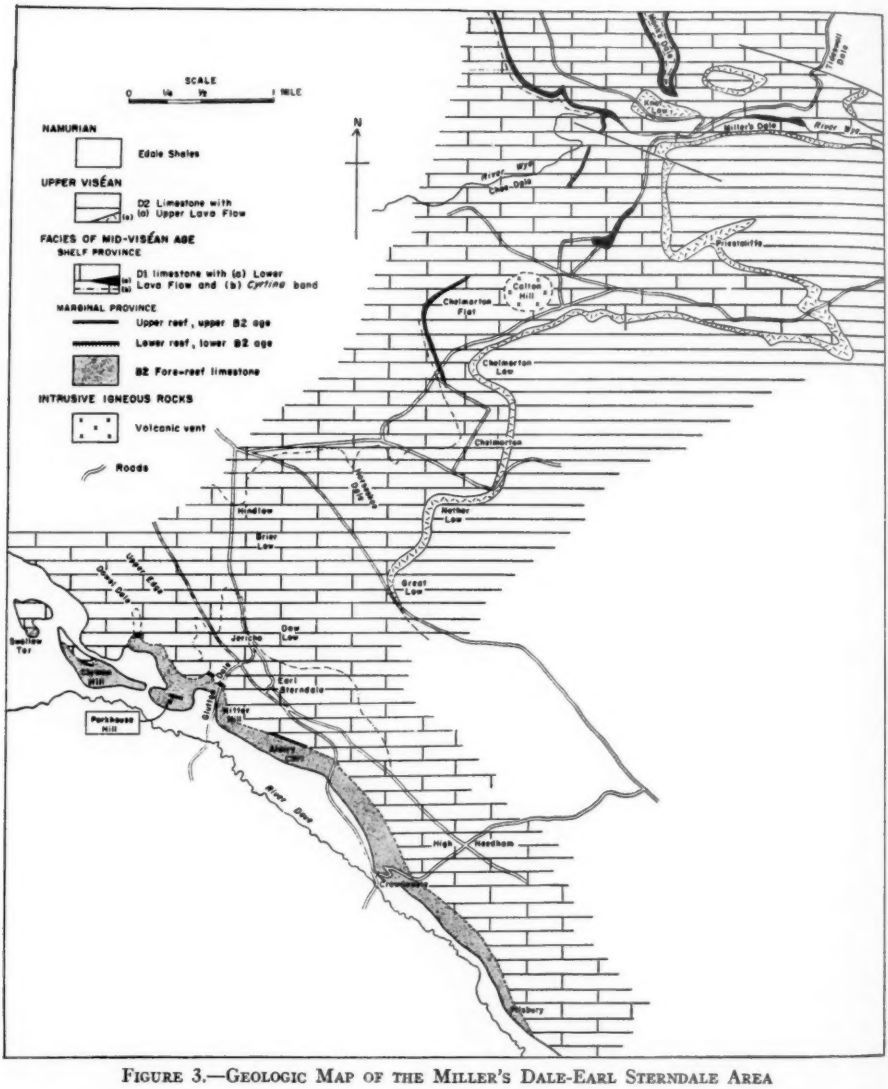


FIGURE 3.—GEOLOGIC MAP OF THE MILLER'S DALE-EARL STERNDALE AREA

both sides of the Winnats Pass, where it is nearly 100 feet high. On the shelf side it passes into back-reef and shelf limestones that are just below the *Cyrtina* band which crops out about 20 feet higher than the top of the reef. On the basin side it passes into steeply dipping fore-reef limestones with a rich brachiopod-mollusk fauna and *Davidsonina septosa* and *Palaeosmilia murchisoni* which indicate a D1 age.

Above the level at which this reef dies out, almost-horizontal shelf limestones pass gradually into reef-complex limestones. They gradually overstep the lower reef and come to rest above steeply dipping fore-reef limestones. The dip in the fore-reef limestones must therefore be depositional.

At higher horizons an upper reef developed, which is exposed for 1200 feet on the summit ridge of Treak Cliff. It is about 30 feet wide

and 50 feet high. To the west it passes into back-reef and shelf limestones of upper D1 age, and to the east into fore-reef limestones that contain upper B2 goniatites.

The section across the reef complex in this area (Fig. 6) is similar to that across the reef complex at Earl Sterndale (Fig. 5).

Reef limestones are poorly exposed southwest of Windy Knoll (Fig. 7). Evidence suggests the reefs developed in the same way as they did elsewhere. The lower reef is exposed in the extreme southern part of the shallow gorge running to Peakshill Farm, between Snels Low and Middle Hill. North of this reef, fore-reef limestones contain lower B2 goniatites and *Palaeosmilia murchisoni* and *Dibunophyllum bourtonense*; to the south are shelf limestones, in which the *Cyrtina* band crops out about 20 feet higher than the top of the reef. A small outcrop believed to represent this lower reef is seen north of the Bull Pit. The upper reef of this area is exposed at the summit of Snels Low. Three small outcrops occur on Middle Hill.

A section across the reef complex in this area (Fig. 7) shows the upper reef almost directly above the lower one. The reasons for this are not known.

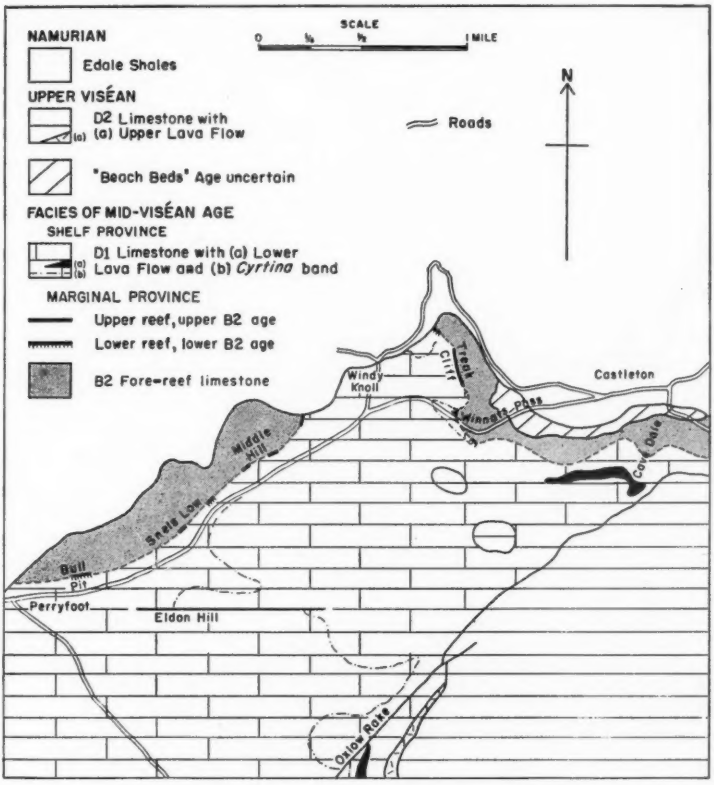


FIGURE 4.—GEOLOGIC MAP OF THE CASTLETON AREA

Reef-Complex Limestones

The reef limestones are built of a rigid framework of *in situ* organic skeletons. The interstices are filled with fine-grained sediment, some of which is detrital. The framework is constructed of stromatolitic algal deposits, with less important contributions from encrusting colonies of *Fistulipora minor* M'Coy and *Chaetetes depressus*; in places *Lithostrotion irregulare* appears to have played a minor part. Lithistid sponges, which are re-

stricted to the reef, may have trapped some loose sediment.

In some places the sediment intercalated between this framework makes up as much as half the rock; much of it is fine-grained and contains no terrigenous detrital minerals. Shell debris is rare. In the fine matrix of many of these limestones, bands of fibrous calcite crystals of secondary origin are present.

The reasons for the dying out of the lower reef are not fully understood. Newell *et al.* (1953, p. 106) pointed out that if "the rate of downwarping exceeds the growth potential of the reef, it will be drowned at comparatively shallow depths". It is possibly significant that the lower reef (and also the upper one) grew steeply upward; this suggests that there might have been an excessive rate of downwarping.

The reef limestones contain many well-preserved fossils. The complete fauna is shown in Table 3. The corals *Dibunophyllum bipartitum*, *D. bourtonense*, and *Palaeosmilia murchisoni* suggest that they are of D1 age.

The back-reef limestones occur just behind the reef and typically are obscurely bedded aphanitic limestones with bands of fibrous calcite crystals. The dip of the limestones, where it can be discerned, is gentle, in conformity with that of the shelf limestones with which they are interbedded and into which

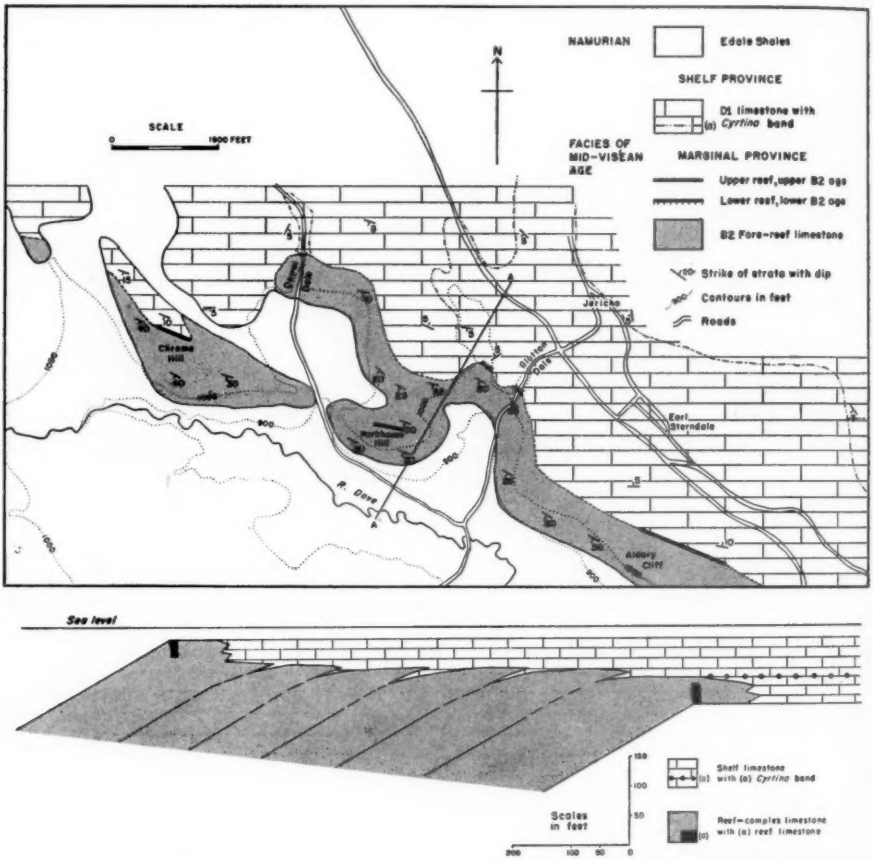


FIGURE 5.—GEOLOGIC MAP AND IDEALIZED SECTION OF PART OF THE EARL STERNDALE REEF COMPLEX

Idealized section is through the reef complex in the Glutton Dale-Parkhouse Hill area, at the time of deposition.

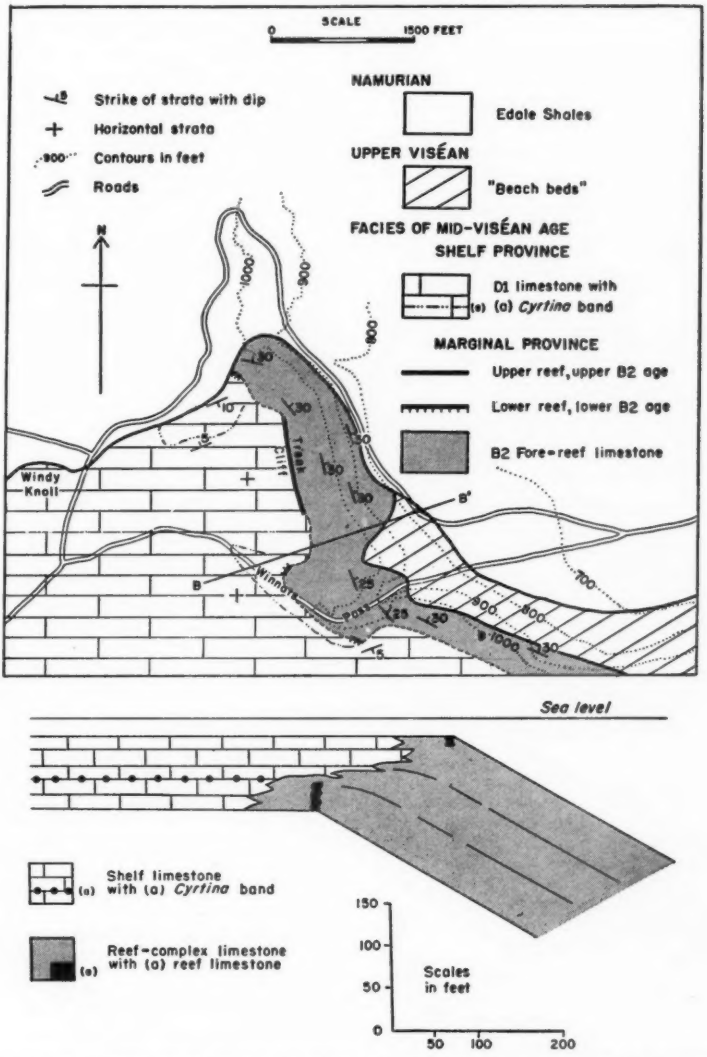


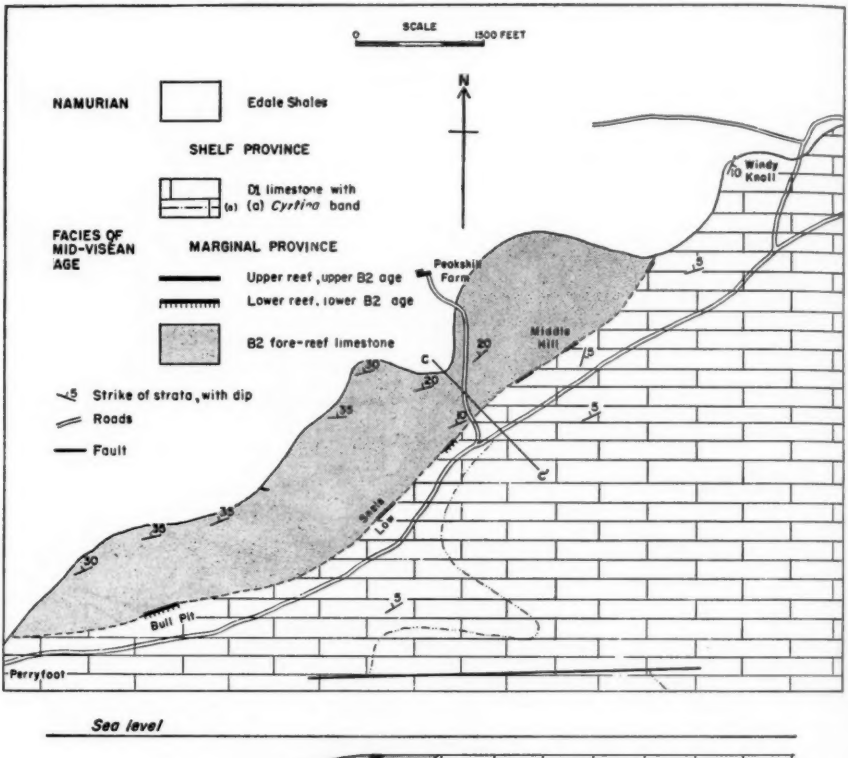
FIGURE 6.—GEOLOGIC MAP AND IDEALIZED SECTION OF THE EASTERN PART OF THE CASTLETON REEF COMPLEX

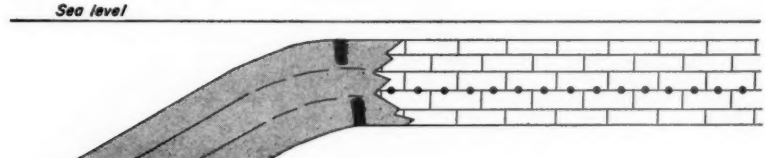
Idealized section is through the reef complex in the Treak Cliff-Winnats Pass area, at the time of deposition.

they pass laterally. They have been recognized only within 200 yards of the reef and generally are best developed within about 50 yards of the reef.

The back-reef limestones are characterized by many reef-phase brachiopods and a few pelecypods, gastropods, and rare goniatites. With these reef-phase fossils are many species characteristic of the shelf limestones.

The fore-reef limestones dip 10°–40° toward the basin with average values of 20°–30°. These may be close to primary dips, although no allowance is made for differential compaction, and the original relief between reef sur-





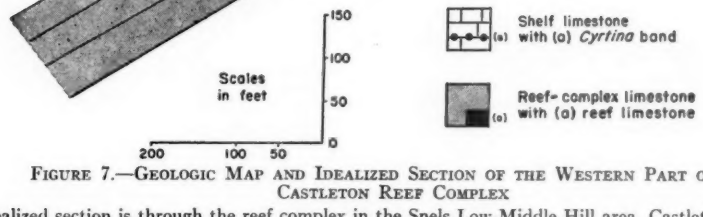


FIGURE 7.—GEOLOGIC MAP AND IDEALIZED SECTION OF THE WESTERN PART OF THE CASTLETON REEF COMPLEX

Idealized section is through the reef complex in the Snels Low-Middle Hill area, Castleton, at the time of deposition.

face and basin floor may have exceeded 400 feet (compare Parkinson, 1953, p. 265). Many of the rocks are massive and unbedded. They are pale-gray, aphanitic limestones very similar to the back-reef limestones. Less frequently there are calcarenites and breccias. Secondary fibrous calcite is common.

These limestones contain an amazing wealth of fossils (Table 3). Mollusks are more common than in other facies, and brachiopods occur in far greater numbers than in the reef (compare Stehli, 1954).

The two subzones of B2 distinguished by Hudson and Cotton (1944) have been recognized. The lower subzone is characterized by *Goniatites* cf. *antiquatus* Bisat, *G. hudsoni*

Bisat, and *G. maximus* Bisat. The upper subzone is characterized by *Beyrichoceras* spp., *G. maximus* var. *b* Bisat, and *Nomismoceras* sp.

The corals *Dibunophyllum bipartitum, D. bourtonense*, and *Palaeosmilia murchisoni* suggest that the rocks belong to the D1 subzone. A very few specimens of *Palaeosmilia regia*, usually regarded as typical of the D2 subzone, have been found at all horizons in these limestones. The suggestion made by Parkinson (1947) that this coral is restricted to the uppermost part of B2, which consequently is of D2 age, can no longer be maintained, and the zonal value of the coral must be questioned.

The presence of *Davidsonina septosa* suggests a D1 age. This has been found only at low horizons, which on field evidence are considered to be equivalent to, or nearly equivalent to, the shelf limestones in which the *Cyrtina* band occurs.

Petrography

Shelf Limestones

Nearly 150 thin sections of these limestones have been prepared from rocks collected over the whole shelf. Except for rocks collected within about a quarter of a mile of the reef, little variation has been observed.

The typical shelf limestones vary only in grain size and in the proportion of the types of organic debris of which they are composed. They are calcarenites with some calcirudites and calcilutites, and are composed of crinoid and brachiopod fragments with minor amounts of foraminifers, coral debris, and ostracods. Algal colonies occur rarely. *Girvanella* spp. occur in encrustations on shell fragments; *Koninckopora inflata*, a dasycladacean alga (Wood, 1942), occurs in about half the slides; *Calcisphaera* sp. is almost ubiquitous; and borings made by algae occur in about 30 per cent of the slides. Crinoid debris, the most important rock former, makes up about half the rock on average and is about 3 times as abundant as brachiopod debris. Foraminifers, corals, ostracods, and algae are quantitatively unimportant, although local concentrations (as much as 10 per cent foraminifers and 15 per cent *Koninckopora*) have been recorded.

Clear calcite cement and minute granular calcite crystals of clay and fine silt size are also present. The origin of these latter is not known. They may be physicochemical precipitates, finely triturated shell debris, or algal dust (Wood, 1941).

Most of the rocks are poorly sorted, differing in this respect from many calcarenites. The local source of much of the material of the rocks is probably partly responsible for the poor sorting, and current action varying somewhat in strength may be indicated.

Terrigenous detrital minerals are absent, although the presence of alumina in insoluble residues (determined spectrographically) may indicate minor amounts of clay minerals. The only noncarbonate mineral visible in thin section is authigenic quartz.

The term spergenite was proposed by Pettijohn (1949, p. 301) for those calcarenites with less than 10 per cent quartz, and from 10 to 80 per cent oölitic grains. Such rocks occur within about a quarter of a mile of the reef. They are composed of shell debris, simple, composite, and oölitic grains set in a clear calcite cement. Clay-sized crystals, which are almost ubiquitous in the coquinas and microcoquinas just described, are almost absent.

Well-rounded shell debris makes up 30–60 per cent of these rocks. Crinoid and brachiopod debris (the former being about twice as common) are predominant, but foraminifers, bryozoans, coral detritus, ostracods, and algal colonies are present. From 25 to 40 per cent of the rocks is made up of simple, composite, and oölitic grains.

The simple grains are about 0.08–0.5 mm in diameter. Most of them are elliptical or circular and are made up of varying proportions of shell debris and of calcite crystals which range from 3 to 7μ; some grains appear to be made up solely of such calcite crystals.

Composite grains are made up of simple grains and shell fragments cemented with calcite crystals identical with those forming the matrix of the simple grains. Rarely oölitic grains are incorporated. Composite grains range from 0.5 to 4 mm in diameter, with an average of 1 mm. Many are irregular, but some are partly rounded.

Oölitic grains are nowhere abundant. Simple and composite grains and shell fragments may show oölitic laminae on their external surface. The diameter of these grains ranges from 0.5 to 1.5 mm, with an average of 1 mm. The thickness of the oölitic laminae is up to 0.2 mm; some individual laminae which are composed of calcite crystals with a radial elongation, are as thin as 0.01 mm.

As the proportion of oölitic grains decreases,

the amount of fine-grade interstitial sediment increases. However, it rarely makes up as much as 5 per cent of the rock.

In contrast to the coquinas and microcoquinas, the spergenites are well sorted, and the shell fragments are well rounded.

True spergenites are confined to within a quarter of a mile of the reef, but limestones containing many simple and composite grains, showing affinities with the spergenites, have been found farther away; in the *Cyrtina* band they occur 2 miles from the reef.

M. Black (1933) and Illing (1954) described the environment in which such rocks could form. Black found that oölitic sediments on the Great Bahama Banks formed only in the marginal areas, and he believed that they represented physicochemical precipitates of calcium carbonate that formed where currents were sufficiently strong to keep the sand-sized nuclei of the oölitic grains in constant motion. This precipitation can occur only in shallow warm waters supersaturated with calcium carbonate. Illing described sediments forming on the Bahama Banks which contained up to 88 per cent of nonskeletal grains, which he divided into "lumps", "grains", "aggregates", and fecal pellets, all of which may act as nuclei for oölitic accretion. Illing's "grains" and "aggregates" seem to be identical with the simple grains of the spergenites, and the "lumps" compare with the composite grains.

It is suggested therefore that the spergenites and limestones with spergenitic affinities formed in the same way as did the calcarenites described by Black and Illing. They must have formed in a similar environment, namely in warm, shallow seas supersaturated with calcium carbonate that were subject to strong current action. The spergenites probably required stronger current action for their formation than did those limestones that do not contain oölitic grains (Illing, 1954). The distribution of these limestones indicates that such conditions were realized only in the marginal parts of the shelf province. In a study of Devonian limestones Beales (1956, p. 869) has reached the same conclusions.

Reef-Complex Limestones

In hand specimens the reef limestones are distinguished from the other limestones of the reef complex by an organic framework constructed of stromatolitic algal deposits. The fine sediment within the framework is imilar to the fine sediment making up the back-reef and fore-reef limestones. The characteristics of the reef-complex limestones have been determined from nearly 180 thin sections.

They contain a variable, generally small, proportion of shell debris, set in a matrix of minute calcite crystals. In many places this matrix is partially recrystallized. As a result, clotted limestones (Cayeux, 1935; Wood, 1941) and bands of fibrous calcite crystals (W. W. Black, 1952, p. 195) have developed.

The shell debris includes fragments of crinoids, brachiopods, mollusks, bryozoans, foraminifers, and ostracods, with small unidentifiable fragments. Crinoids do not predominate, in contrast to the shelf limestones. Algae are represented by *Girvanella* colonies, rare *Koninckopora*, *Calcisphaera*, and occasional borings in shell fragments; in many slides *Girvanella* tubules can be faintly discerned in the matrix. Modal analyses of three back-reef limestones showed an average shell content of 10 per cent; in three reef limestones, the average was 25 per cent, and four fore-reef slides averaged 14 per cent (compare with W. W. Black, 1954, p. 278); as little as 2 per cent may be found.

These limestones are distinguished by their fine matrix, which is a granular mosaic of interlocking calcite crystals 3–8 μ in size. In many cases this matrix is partially recrystallized, but otherwise it is structureless.

Certain deposits of crystalline fibrous calcite in the reef limestones are considered to represent precipitated infillings of primary cavities within the organic framework (Newell *et al.*, 1953, p. 107; Newell, 1955a). The cavities may be 2 inches across and are commonly filled with calcite crystals that are either anhedral or elongate perpendicular to the surface. Some of these cavities have been lined with fibrous calcite crystals and then filled with fine detrital sediment. The calcite must therefore have been precipitated early in the history of the rock, when the cavity was still open to the access of fine sediment, as suggested by Cullis (1904) and Newell (1955a).

Origin of the Reef-Complex Limestones

Microcrystalline limestones, characteristic of reefs, occur in Silurian reefs of Gotland (Hadding, 1941, 1950), Upper Carboniferous reefs of New Mexico (Plumley and Graves, 1953), Permian reefs of Texas and Greenland (Newell *et al.*, 1953; Newell, 1955b), Cretaceous and Tertiary reefs of the Middle East (Henson, 1950), and many others.

Walther (1885) concluded from his study of Tertiary algal limestones of Syracuse (Sicily) that recrystallization of algal limestones could result in rocks composed of crystals smaller than those in limestones of clastic origin. Hadding (1950), Newell *et al.* (1953), and W. W. Black (1954) suggested that some of the microcrystalline limestones of reefs represent such recrystallized algal limestones.

Aphanitic limestones may originate in at least three other ways: by bacteriological precipitation; by physicochemical precipitation; and by the trituration of shell debris.

Any theory which attempts to explain the genesis of the calcilutites in the Derbyshire reef complex must be reconcilable with the following facts:

(1) That aphanitic limestones with little or no shell debris formed in the vicinity of areas where clastic limestones composed for the most part of shell debris accumulated;

(2) That many of these aphanitic limestones are unbedded and at best are poorly bedded;

(3) That they accumulated on 15°–30° slopes;

(4) That they accumulated in areas where, from evidence provided from breccias and encrusting growth forms, there was considerable wave and current action.

Bavendamm (1932), reviewing the literature on the precipitation of calcium carbonate by bacteriological action, concluded that such precipitation would not occur in open sea water, as the bacteria concentration is too small to bring about the necessary changes; precipitation would be limited to mangrove swamps. It therefore does not appear to be a likely mode of origin for the rocks under consideration.

The most favorable environment for the production of physicochemically precipitated muds of calcium carbonate is warm, shallow, relatively sluggish water, with greater salinity than normal (M. Black, 1933; Smith, 1940). Such conditions might have prevailed at times in the center of the shelf province, but it seems unlikely that all of them were realized in the marginal province.

Calcilutites in the shelf province represent finely triturated shell debris and a little material of unknown origin. These, however, differ significantly from the aphanitic limestones of the reef complex in containing abundant recognizable shell debris.

Sediments formed in any of these three ways would accumulate as soft, unconsolidated lime oozes. It is difficult to reconcile such a mode of origin with strong current action. Kuenen (1935) has shown that very mobile muds can accumulate on slopes of 15° in thicknesses up to 3 feet, but this is probably the maximum slope on which sedimentation can take place. It seems unlikely that an appreciable thickness of sediment can develop on such a slope.

If the aphanitic limestones formed in any of these ways, it is remarkable that they contain so little recognizable shell debris; many present-day examples of physicochemically precipitated muds contain up to about 70 per cent shell debris (M. Black, 1933).

The only remaining hypothesis considers them as partly representing recrystallized algal deposits. Positive evidence is lacking, and the only support for this suggestion lies in the *Girvanella* tubules which are common in the fine matrix of the limestones. W. W. Black (1954, p. 292) concluded from a study of Lower Carboniferous reefs of northern England that aphanitic limestone in them had probably formed from the recrystallization of algal deposits.

It is believed that the fore-reef slope had a relief of about 400 feet. Since nearly all marine plants are confined to depths of less than 300 feet, there is a lower limit on this slope below which the limestones cannot have formed by the recrystallization of *in situ* algal deposits. Limestones at several localities at low levels on the fore-reef slope are characterized by an abundance of crinoid remains that have every appearance of being near their place of growth. It is probable that the crinoids here are becoming important rock formers, once more assuming the role they fulfilled in the shelf seas.

Paleontology

General

More than 3000 fossils from the various limestones have been studied in the laboratory, and a great many more have been examined in the field. From the observed distribution, an attempt has been made to reconstruct the life distribution (Fig. 8). The basin facies has not been considered.

Corals

Simple and compound corals have been found throughout the shelf limestones; the same genera and species occur over the whole shelf province. Many of the simple corals are fragmented, with the dissepimental areas lost, probably by erosion. Some of the compound

corals have been overturned by wave or current action.

In the reef limestones some large colonies of *Lithostrotion irregulare* are in growth position, and small encrusting colonies of *Chaetetes de-*

Brachiopods

The shelf limestones of Miller's Dale accumulated about 5 miles from the reef, and the brachiopod fauna is probably uncontaminated

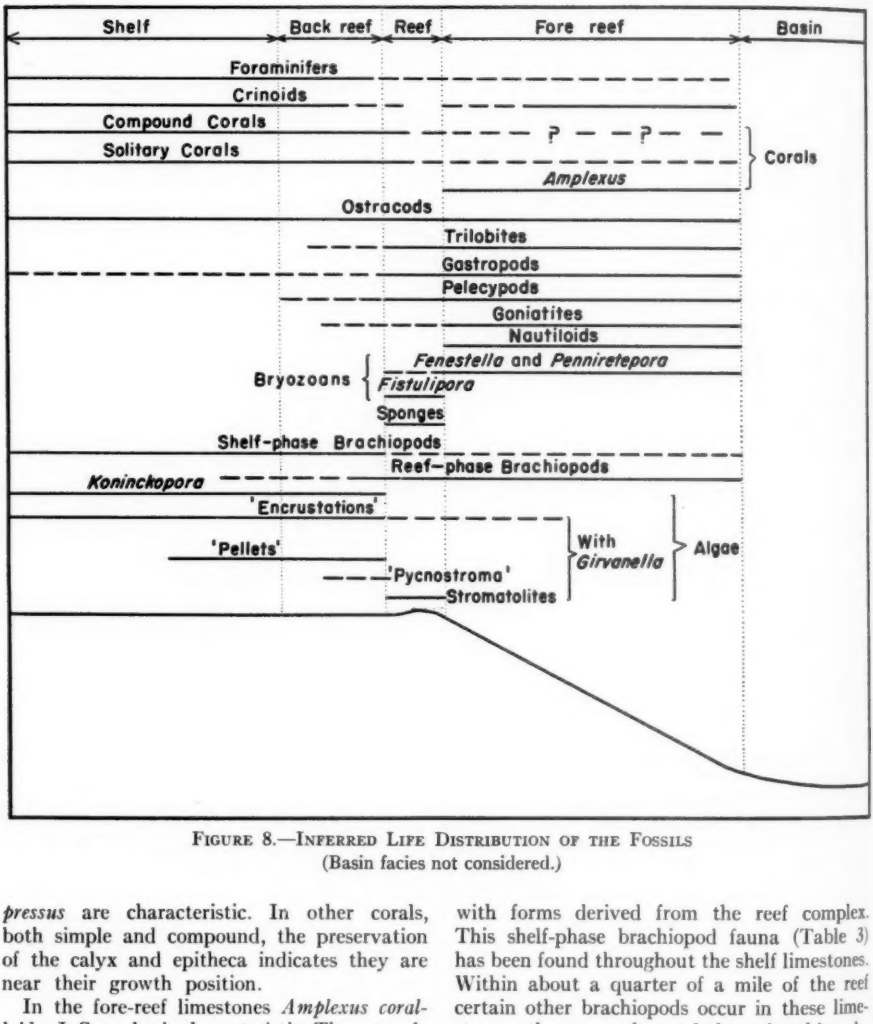


FIGURE 8.—INFERRED LIFE DISTRIBUTION OF THE FOSSILS
(Basin facies not considered.)

pressus are characteristic. In other corals, both simple and compound, the preservation of the calyx and epitheca indicates they are near their growth position.

In the fore-reef limestones *Amplexus coralloides* J. Sowerby is characteristic. Those corals typical of the shelf limestones are rare, and, although some of the simple forms appear to be *in situ*, none of the compound corals can be shown to be in growth position.

The suggested life distribution is in close agreement with what is already known about the facies distribution of Lower Carboniferous corals (Hill, 1938).

with forms derived from the reef complex. This shelf-phase brachiopod fauna (Table 3) has been found throughout the shelf limestones. Within about a quarter of a mile of the reef certain other brachiopods occur in these limestones; these are the reef-phase brachiopods. The number of individuals, genera, and species of these increases toward the reef.

Most of the shelf-phase brachiopods have been found in the reef complex, but they are far less common there. None of them were found at more than half the localities from which brachiopods have been collected in the fore-reef limestones.

In all areas most of the shelf-phase brachiopods occur as single valves, showing that they have been moved from their life position. Some complete individuals were found in the reef complex; a few of them probably lived there, but many others may have drifted. They were certainly a rare element of the reef-complex fauna.

A rich brachiopod fauna occurs in the reef complex (Table 3). Many more individuals and also more species occur in the fore-reef than in the reef and back-reef limestones. Since most of these specimens are complete shells, and many have delicate spines still attached to the shell, it seems certain that they have not been moved far from where they lived. Most species had a considerable depth range (compare Stehli, 1954).

Many of the reef-phase brachiopods in the marginal shelf limestones are found as single valves, some of which may be derived from the reef complex. Since the valves are not fragmented, and, since some spines are still attached, it is quite possible that they are still near the place where they lived.

Pelecypods

In the shelf and back-reef limestones, pelecypods occur only within about 200 yards of the reef, most commonly in microcrystalline back-reef limestones. They are rare. Single valves of byssiferous and burrowing forms are present.

Pelecypods are more common in the reef limestones where more species occur. As in the shelf limestones single valves dominate. Byssiferous, burrowing, free-living forms and a possible rock borer (*Lithodomus lingualis* Phillips) occur.

Pelecypods are most abundant in the fore-reef limestones, where the greatest number of genera and species occur. Entire shells are more common here. Byssiferous, burrowing, free-living, and (?) boring forms are found. The majority are byssiferous.

Since in many pelecypods the dentition is poorly developed, a large preponderance of single valves might be expected, and would not necessarily indicate that the shells have travelled far. Entire shells, however, are almost certainly near their life position. Complete single valves which retain their surface ornament are also probably near where they lived. On this basis the distribution shown in Figure 8 has been deduced.

A similar distribution was found by Newell (1955b) in Permian reefs of Greenland, but Newell *et al.* (1953, p. 203) found a somewhat different one in Permian reefs of Texas.

Gastropods

At distances of more than 200 yards from the reef, *Bellerophon* is the only gastropod found in the shelf limestones, and it occurs over the whole shelf. In strata within 200 yards of the reef, *Bellerophon*, *Euomphalus*, *Naticopsis*, and *Straparollus* occur. In the reef limestones gastropods are rare, and few species occur. Gastropods are most common in the fore-reef limestones, where the greatest number of genera and species occur. Both low- and high-spired forms have been found at all levels on the fore-reef slope.

Since in all areas the gastropod fauna consists of well-preserved individuals on which surface ornament is still intact, it is considered that the present distribution corresponds to the life distribution.

This distribution contrasts with that in Permian reefs of Texas (Newell *et al.*, 1953, p. 203), where gastropods are most common in shelf strata just behind the reef.

Nautiloids

Only two indeterminate coiled forms have been found in shelf limestones, and both came from within 50 yards of the reef. In the reef two specimens of *Orthoceras* have been found. Nautiloids are most common—although far from abundant—in the fore-reef limestones, where there are both straight and coiled types. Rare lenses of fore-reef limestone contain considerable numbers of species and individuals, as in Dowel Dale, Earl Sterndale.

Since modern nautiloids are free swimming, and since they are often transported far from their life habitat, it is far from certain that the above distribution parallels the life distribution. Flower (1942) considers that Paleozoic nautiloids were largely facies organisms. The above distribution appears to support this statement, for the evidence is most reasonably interpreted by considering that during life the nautiloids were confined to the fore-reef facies.

Goniatites

These have a distribution similar to the nautiloids. They are most common in the fore-reef facies, in which they are typically found in pockets crowded with well-preserved specimens of all sizes. In the reef only one specimen has been found. In back-reef limestones at

Earl Sterndale, a pocket of lower B2 goniatites has been found; the specimens occur in all sizes of well-preserved individuals, indicating that it is not a drifted assemblage. Hudson and Cotton (1944), however, regard such assemblages as drifted ones. Goniatites are rare in the marginal shelf and back-reef limestones and are not found far from the reef.

It is suggested that this distribution parallels the life distribution.

Sponges

Sponges are essentially restricted to the reef limestones. They are a common characteristic element of the reef fauna. Most specimens are complete. In Permian reefs of Texas (Newell *et al.*, 1953, p. 202) sponges are confined to the reef.

Trilobites

Shrock and Twenhofel (1953) consider that isolated pygidia, thoraxes, and cephalons in general represent moult casts. These have a low specific gravity and a high surface area relative to volume and can therefore be transported long distances by slowly moving waters of low competency. Entire shells indicate that the specimen has been buried *in situ*.

Trilobites are most common in the reef and fore-reef limestones; entire specimens have been found in both facies. In the shelf and back-reef limestones a few isolated cephalons and pygidia have been found within about 200 yards of the reef. These may have drifted, but they are not fragmented or rounded by current action, and they may have lived where they are found.

Ostracods

Hand specimens have been found in all the limestones but are most common and seem to be larger in the reef and fore-reef limestones. They occur in 60–70 per cent of the thin sections of all facies. They probably lived in all facies, apparently showing a slight preference for the reef and fore reef.

Bryozoans

In back-reef and shelf limestones macroscopic specimens are rare and are limited to a few fragments which are most common near the reef and absent at distances greater than 400 yards from the reef. In the reef, entire specimens of *Fenestella*, *Fistulipora* and *Pennireteporа* occur; *Fistulipora* was found at all outcrops. *Fenestella* and *Pennireteporа* are common in the fore-reef facies. The distribution of bryozoan debris in thin sections is shown in Table 2.

During life, bryozoans were probably restricted to the reef and fore-reef facies; *Fenestella* and *Pennireteporа* lived in both, but *Fistulipora* was confined to the reef. An almost identical bryozoan distribution occurs in Permian reefs of Texas (Newell *et al.*, 1953, p. 112–113).

Crinoids

Throughout the shelf limestones, stem sections, scattered columnals, and head plates are present in great amounts, and in most thin sections crinoid debris makes up about half the rock and is the most important rock former. In the back-reef aphanitic limestones, crinoids are rare.

Crinoid debris is rare in the reef and is confined to small stem fragments and scattered plates.

In the aphanitic limestones at high levels on the fore-reef slope, crinoids are rare, and only stem fragments and isolated plates have been found. At several localities low on this slope, however, some limestones are made up almost entirely of large, long crinoid stems and some complete heads. These have every appearance of being near where they lived.

In those reef-complex limestones with little macroscopic debris, it is unusual to find as much as 5 per cent crinoid debris in thin section.

In all facies crinoids have been broken and dispersed by wave and current action. The abundance of crinoids throughout the shelf limestones indicates that they lived in considerable numbers throughout the shelf province. Their general absence from the reef limestones suggests that they were a rare element of the reef fauna, if in fact they lived there, and they appear to have been sporadic in the back-reef facies and on the upper levels of the fore-reef slope. At lower levels on this slope they were apparently common.

Foraminifers

In some thin sections of the shelf limestones foraminifers constitute 1–5 per cent of the rock. In the aphanitic limestones of the reef complex they are far less common, and they

are rare or absent in about 70 per cent of the thin sections from the fore-reef facies. Associated with the stromatolites of the reef is *Aphralysia carbonaria* Garwood, which is thought to be an encrusting foraminifer comparable with *Nubecularia*.

it is most abundant in the reef and fore reef; shelfward its abundance falls to an almost constant value.

The stromatolites constructed the framework of the reef, and *in situ* specimens are restricted to the reef, with a little debris in neighboring

TABLE 2.—DISTRIBUTION OF BRYOZOANS, GIRVANELLA, AND KONINCKOPORA IN THIN SECTIONS

Facies	Number of slides	Percentage of slides with bryozoans	Percentage of slides with *Girvanella*	Percentage of slides with *Koninckopora*
Fore reef	104	73	67	7
Reef	38	61	95	4
Shelf:				
a. Up to ¼ mile from reef	102	37	39	38
b. ¼–1½ miles from reef	24	4	16	50
c. 1½–2 miles from reef	10	0	20	50
d. 2½–3 miles from reef	9	0	11	55
e. 4–5 miles from reef	38	0	8	53

This limited survey suggests that foraminifers were most common in the shelf seas. Evidence suggests that a possible encrusting foraminifer was restricted to the reef.

Algal Colonies Containing Girvanella

Girvanella occurs in a variety of algal colonies, namely:

(1) The stromatolites
(2) *Pycnostroma*-like forms. These are pisolitic forms with a diameter up to 1 mm. In thin sections they show lamination and contain *G. ducii* Wethered.
(3) "Pellets". These are sand-sized grains of algal filaments twisted into a spaghetti-like mass. *G. ducii*, *G. incrustans* Wethered, and *G. staminea* Garwood occur; some have more than one species in a single pellet.
(4) Encrustations on shells. These are thin coatings built up of intertwined algal threads. *G. ducii*, *G. incrustans*, *G. staminea*, and *Ortonella tenuissima* Garwood occur.
(5) "Crusts". A single specimen found in shelf limestones near Chelmorton Flat was built up of thin horizontal laminae. *G.* sp. was present.
(6) *Ottonosia*-type colonies. A single specimen was found in fore-reef limestone at Cave Dale, Castleton. It was a laminated spherical colony with a diameter of 2 cm and contained *G. staminea*.

The general distribution of the genus in thin sections is seen in Table 2, which shows that

limestones. The distribution of this type of algal colony in thin section is seen in Figure 9. These colonies were probably restricted during life to the reef. Newell *et al.* (1953, p. 112) showed that similar colonies were responsible in part for building Permian reefs of Texas.

The *Pycnostroma*-like forms are rare and have been found only in back-reef strata within about 50 yards of the reef.

"Pellets" are most characteristic of the spergenites and limestones with spergenitic affinities and, like the limestones in which they are found, are restricted to strata within about a quarter of a mile of the reef. Their distribution in thin section is seen in Figure 9; it probably corresponds to the life distribution.

Encrustations on shells are not common but seem most characteristic of the shelf limestones and occur over the whole shelf (Fig. 9).

Koninckopora

Hand specimens of this alga are rare in the fore reef and have not been found in the reef limestones. They occur throughout the shelf limestones, and, although not common, at certain localities almost-complete specimens occur in vast numbers, which probably represent original thickets of the plant (Wood, 1942).

The distribution of the plant in thin section is seen in Table 2. In the shelf limestones *Koninckopora* may make up about 1–3 per cent of a thin section and up to 15 per cent.

A high percentage is found in many spergenites and limestones with spergenitic affinities, but the present work does not entirely support Wood's conclusion (1942) that the plant is invariably associated with oölitic sediments, as high values occur in nonoölitic sediments.

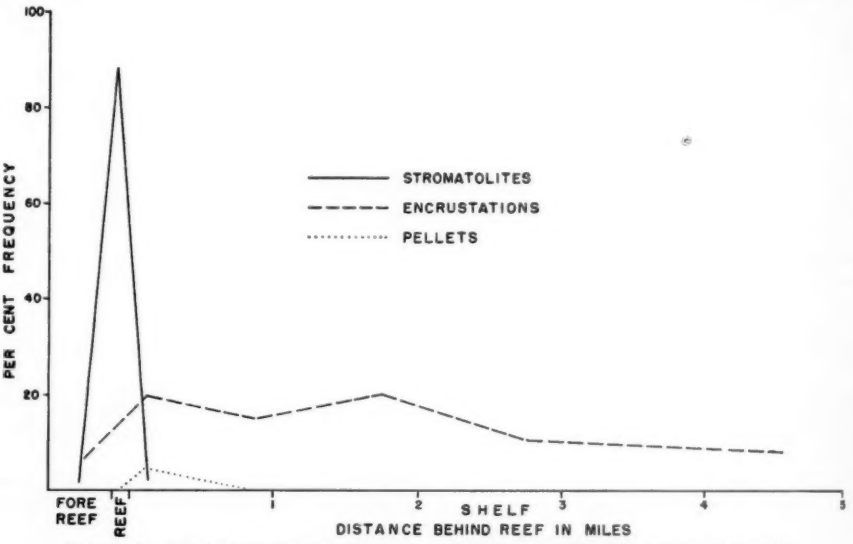


FIGURE 9.—THIN-SECTION DISTRIBUTION OF ALGAL COLONIES CONTAINING *Girvanella*

The plant was probably restricted to the shelf seas, and the few fragments in the reef and fore reef represent drifted specimens. This conclusion is supported by Wood (1942). A similar distribution of dasycladacean algae was found in Permian reefs of Texas (Newell *et al.*, 1953, p. 195).

PALEOECOLOGY

Submarine Topography

Abundant evidence indicates that the shelf and reef limestones were deposited in shallow water. In the reef the abundance of calcareous algae and their brecciation is indicative of shallow, wave-disturbed waters. Newell *et al.* (1953, p. 202) consider that optimum depth conditions for the type of algae found in these reefs "was considerably less than thirty meters." In the shelf limestones the presence of dasycladacean algae—recent forms of which are most abundant at depths of 9–15 feet and do not occur below 90 feet (Cloud, 1952)—and of spergenites that contain grains that at the present form in waters less than about 30 feet deep, indicate shallow water. The random orientation of shell debris, the presence of overturned and fragmented corals, the fragmented brachiopod shells, and the sorting of crinoid debris indicate deposition in water subject to strong current action, suggestive of shallow depth.

In several areas, horizontal shelf limestones rest directly above steeply dipping fore-reef limestones, showing that this dip is depositional. The shallow shelf province was therefore surrounded by a deeper-water area in which the basin facies was deposited. The relief along this fore-reef slope was at least 400 feet.

Hydrography

Many hydrographic features of the area at the time of deposition are unknown, but it is possible to make suggestions as to the probable temperature, salinity, gas content, and strength of current action of the waters over the shelf and marginal provinces.

The chemistry of calcium metabolism requires that fossil reefs, like present-day ones, should have formed in agitated, warm, clear, shallow seas (Vaughan, 1911). Sverdrup *et al.* (1942) point out that deposits high in calcareous skeletal structures are generally restricted to areas where the ionic product $[Ca^{++}] \times [CO_3^{=}]$ is large, and Revelle (1934) showed that this product was raised by in-

crease in salinity and temperature and decrease in the total carbon dioxide content.

The salinity effect is small. Precipitation is most likely to occur in areas of high temperature and active photosynthesis. Temperatures were probably high over the shelf, since precipitated material occurs in the spergenites. The presence of dasycladacean algae in the shelf limestones provides some confirmation, since these algae at the present day are limited to tropical or subtropical seas.

The increase in depth down the fore-reef slope must have been accompanied by a fall in temperature, but the amount is not known.

Certain stenohaline animal groups, such as the corals and echinoderms, are present in both provinces and suggest normal salinity. The physicochemically precipitated calcium carbonate in the spergenites, however, probably indicates some slight supersaturation of the water with calcium carbonate over the parts of the shelf in which these rocks formed.

The very light color of the limestones indicates near absence of organic matter. The organic matter must have been oxidized, either on the sea floor or during the early stages of diagenesis. Oxygen must therefore have been present in appreciable amounts in the sea water, and high temperatures to aid the oxidation are suggested. The upper layers of the sediment probably had a positive Eh (oxidation-reduction potential). Since Eh decreases with increase in depth in the sediments (Zobell, 1946), the oxidation must have been a very early diagenetic modification. The general absence of pyrites from the limestones supports the hypothesis.

The physicochemical precipitates in the spergenites suggest that the pH of the sea water was as high as 8.0, for at lower pH values the carbonate ion is repressed into bicarbonate ion and carbonic acid (Krumbein and Garrels, 1952), and precipitation in acid environments is impossible. The pH is strongly affected by the carbon dioxide content; the carbon dioxide content was probably low, because of agitation, high temperatures, and photosynthetic activity.

The brecciation within the reef limestones, the occurrence of reef-derived material in adjacent limestones, and the encrusting growth forms of the reef builders suggest that the reef was subject to considerable wave and current action. Vaughan (1911) considers that strong current action is characteristic of reefs, both ancient and modern. Since the reef developed at the edge of a topographic high, strong current action would be anticipated (Sverdrup *et al.*, 1942).

Dietz and Menard (1951) give the maximum depth of wave action as 36 feet, but to what depth current action extended below this on the fore-reef slope is unknown. Available evidence suggests that slight current action extended to low levels on the slope.

Over the shelf, current action is believed to have been variable. Thus overturned compound corals, single valves of brachiopods, and fragments of corals, brachiopods, and crinoids indicate strong current action. The spergenites formed under appreciable current action. In contrast, the bedded calcilutites could form only when currents were negligible. It is thought, by analogy with modern shelf seas, that for long periods of time the waters over the shelf were undisturbed by currents.

Major Biotopes

A biotope is defined by Hesse *et al.* (1951) as an area showing uniformity in the principal habitat conditions; the association of animals inhabiting such an area is a biocoenosis. In the present instance five biotopes can be recognized (Fig. 10).

In the shelf biotope, the fauna was dominated by crinoids. Simple and compound corals and a few species of brachiopods are common. Foraminifers occur; ostracods and gastropods are rare. The flora includes encrustations of *Girvanella*, the dasycladacean alga *Koninckopora*, *Calcisphaera*, and rare penetrative algae.

The sediments are calcarenites with some calcirudites and calcilutites. In the marginal areas spergenites and limestones with spergenitic affinities occur. Detrital terrigenous minerals are absent, as in other biotopes.

It is believed that the shelf was an area of warm shallow waters, generally of normal salinity but supersaturated with calcium carbonate at times. They were well oxygenated and alkaline, and it is suggested that they were subject to wave and current action varying considerably in strength; in general the waters were probably calm.

In the back-reef biotope, crinoids are common in places. Many reef-phase brachiopods occur together with some shelf-phase forms. A few mollusks—gastropods, pelecypods, and rare goniatites,—are present, with simple and compound corals, foraminifers, ostracods, and a few trilobites. The flora comprises encrustations of *Girvanella*, *Pycnostroma*-like forms, *Koninckopora*, and *Calcisphaera*. It has been suggested that some of the aphanitic limestones of this facies represent recrystallized algal deposits.

The rocks are unbedded aphanitic limestones

which pass into, and are interbedded with, normal shelf limestones.

The environment was probably similar to that of the shelf biotope, but, since current action increases toward the margins of modern shelves, currents were probably stronger than over central parts of the shelf.

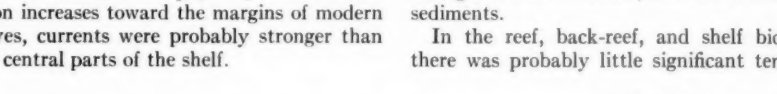


FIGURE 10.—MAJOR BIOTOPES

A, shelf biotope; *B*, back-reef biotope; *C*, reef biotope; *D*, fore-reef biotope; and *E*, basin biotope

The reef biotope is characterized by the stromatolites, tabulate corals, fistuliporoid bryozoans, and lithistid sponges. Reef-phase brachiopods are common, and pelecypods occur. Gastropods and goniatites are rare. Crinoids are almost absent, and corals are rare. Foraminifers, ostracods, and trilobites occur.

The rocks are unbedded aphanitic limestones which consist of an algal framework, with fine material, some of which is detrital.

The reefs formed in shallow water that was subject to strong wave and current action. It is believed that they developed in well-oxygenated, clear waters of normal salinity.

The fore-reef biotope is characterized by the greatest variety of animal life. Reef-phase brachiopods are abundant, and mollusks, including pelecypods, gastropods, and cephalopods, are best represented here. Crinoids are common at low levels on the fore-reef slope. Trilobites and ostracods occur; foraminifers are rare. Corals are rare, but *Amplexus* is typical. Algae seem to be rare, but some of the rocks may be recrystallized algal deposits.

Most of the rocks are massive aphanitic limestones, but lenses and pockets of detrital limestones made up of shell debris, and rare breccias occur.

The environment must have varied owing to the original slope, at the top of which conditions must have been similar to those of the reef. Lower down the slope the waters must have been calmer and cooler. They were probably well oxygenated and of normal salinity.

Theoretical Considerations

It remains to try to recognize the factors responsible for the differences in species density (population density is unknown) between the biotopes. The environmental factors which appear to vary most are: temperature, depth, strength of current action, and nature of bottom sediments.

In the reef, back-reef, and shelf biotopes, there was probably little significant temperature variation, but down the fore-reef slope there must have been a fall in temperature. By analogy with recent corals, many of which are stenothermal, it is tentatively suggested that the compound corals were confined to the shallow-water biotopes, in part at least, by temperature control. This same control might have been partly responsible for restricting *Koninckopora* to shallow water, for recent members of this family of algae are confined to warm seas.

The increase in depth down the fore-reef slope must have exercised considerable control over the development of algae. Thus *Koninckopora* is limited to the shallow-water biotopes, because green algae require light intensities found only in very shallow water. The stromatolites are likewise confined to the shallow waters of the reef; Newell *et al.* (1953, p. 202) suggest that optimum depths for such algae would be considerably less than 30 m. In conjunction with depth, wave action was probably important in controlling the distribution of these algae. Dasycladacean algae prefer a calm-water environment, a possible reason why *Koninckopora* did not live in the exposed shallow waters of the reef. The stromatolites must have occupied a very specialized habitat, as indicated by their restricted distribution. It seems possible that, like present-day reef building Lithothamnionaceae, they were restricted to areas where constantly agitated water kept their living surfaces clear of silt.

Strength of current action was probably responsible for the encrusting growth forms of the reef builders, and possibly it was responsible for limiting the crinoids to the generally calmer waters of the shelf and low levels of the fore-reef front.

The nature of the bottom sediments seems to be an important factor governing the distribution of many benthonic animals. The sediments of the shelf, by analogy with comparable recent sediments, were unconsolidated and unstable at the time of formation. The aphanitic limestones of the reef complex would appear to represent lithified muds, but evidence against this belief has been presented, and it was suggested that they are partly recrystallized algal deposits, implying the presence of an originally firm substratum. This is in keeping with the rich benthonic fauna. The importance of a hard substratum is exemplified by the rich fauna on the submarine banks in the Bay of Naples in contrast to that found on the surrounding mud-covered sea floor (Walther, 1885).

For such a rich mass of phytoplankton to develop, a rich supply of nutrient salts and organic foods in the euphotic zone would be a necessity (Sverdrup *et al.*, 1942). A rich supply of nutrients was possibly ensured by upwelling of deeper basin waters rich in such salts, somewhat as postulated by Newell *et al.* (1953, p. 199) and Gardiner (1931).

Algae in Other Lower Carboniferous Reefs

Reef limestones, identical with those in Derbyshire, in the Craven reef complex, Yorkshire, prove that the development of algae in the Derbyshire reef complex is not merely a local phenomenon.

The Craven reef complex has been examined at Settle, Malham, and Cracoe, and reef limestones were recognized on the southern flanks of the High Hills, Settle; on Cawden Hill, Malham; and on Elbolton Hill, near Cracoe. As in Derbyshire these reef limestones are composed of algal deposits, with encrusting colonies of fistuliporoid bryozoans and tabulate corals. Lithistid sponges are common.

A search for reef limestones in the reef knolls of *C* age in the Bowland area, Lancashire, was unsuccessful, and there is little doubt that they are absent. The author is therefore not prepared to accept W. W. Black's (1954, p. 287) statement that "(lithologically and faunally) the reef limestones of the Cracoe, Malham and Settle areas are identical with those of . . . Bowland."

Stromatolites of the Reef

The stromatolites are referable to the form-genera *Collenia* and *Cryptozoon*, which, as shown by Anderson (1950, p. 6), are not true biologic genera since they usually contain more than one genus of algae. Such generic names are here used in the descriptive sense with no biologic implications.

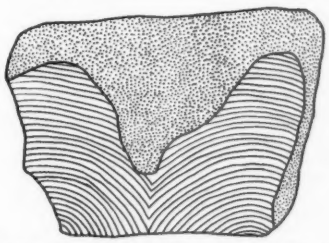


Figure 11.—*Collenia*-Type Algal Colony
Actual size. Upper reef, Aldery Cliff, Earl Sterndale.

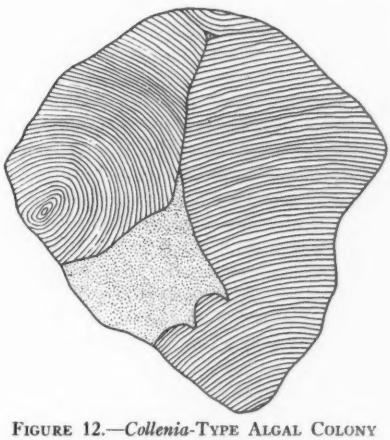


Figure 12.—*Collenia*-Type Algal Colony
×¾. Lower reef, Bull Pit, Castleton.

The algae may be grouped into a series in which the macroscopic structure increases in complexity. Six types can be recognized, but they are not sharply delineated. These are:

(1) Sheetlike forms with horizontal laminae
(2) Broad-based colonies with slightly domed laminae
(3) Colonies with hemispherical dome-shaped heads
(4) Types that put up columns with an approximately circular cross section
(5) Forms that put up conical heads
(6) Complex types with hemispherical dome-shaped heads.

The simplest type is composed of sheetlike growths built up of thin horizontal laminae. This type is not common.

More complex *Collenia* type colonies consist of broad-based, low-growing forms with slightly domed laminae that form undulating crusts (Fig. 11).

The doming of the heads gradually increases until they become hemispherical. Some such heads develop from horizontal laminae, and they are then of *Collenia* type, but more commonly they are associated with contorted laminae and are of *Cryptozoon* type.

Other growth forms that approximate the *Collenia* type are illustrated in Figure 12. These commence as flat sheets in which low domes develop. The continuous sheet gives way to the development of separate heads which have a columnar form with a circular cross section. Still other examples of *Collenia* type put up conical heads and approximate the growth form *Codonophycus*.

The most complex forms are *Cryptozoon* type, in which the laminae are contorted in a most intricate manner, and hemispherical and spherical heads develop to produce a cauliflowerlike growth. These are the most common types.

The algae are built up of successive laminae, which give the structures a characteristic banding. The laminae average from 10 to 20 to the centimeter. The hemispherical dome-shaped heads range from 2 to 4 cm in diameter and from 1 to 3 cm in height; in the spherical heads the diameters range from 2 to 4 cm. In those *Collenia* types which put up columnar pillars of growth, the height of the pillars ranges from 3 to 6 cm, and the diameter from 3 to 5 cm.

In thin section the most obvious feature is the lamination. Recrystallization has destroyed much of the organic structure, so that the rock is composed of very small interlocking granular calcite crystals of clay and silt size.

Girvanella nicholsoni (Wethered), *G. ducii* Wethered, *G. incrustans* Wethered, *G. staminea* Garwood, and *Ortonella kershopensis* Garwood have been recognized, but, owing to the extensive recrystallization, their detailed arrangement within the colonies is unknown. Many slides contain *Aphralysia carbonaria* Garwood, which is thought to be an encrusting foraminifer somewhat similar to *Nubecularia*.

Cryptozoons described by Anderson (1950) from Carboniferous rocks of northern England and southern Scotland, and by Johnson (1940; 1946) from the Pennsylvanian and Permian of Colorado and Kansas, have somewhat similar growth forms. They differ slightly in the size and shape, the thickness of growth laminae, and in the constituent algal species.

References Cited

Anderson, F. W., 1950, Some reef-building calcareous algae from the Carboniferous rocks of northern England and southern Scotland: Yorkshire Geol. Soc. Proc., v. 28, p. 5–28

Bavendamm, W., 1932, Die mikrobiologische Kalkfallung in der tropischen See: Arkiv. f. Mikrobiologie, v. 3, p. 205–276

Beales, F. W., 1956, Conditions of deposition of Palliser (Devonian) Limestone of southwestern Alberta: Am. Assoc. Petroleum Geologists Bull., v. 40, p. 848–870

Black, M., 1933, The precipitation of calcium carbonate on the Great Bahama Bank: Geol. Mag., v. 70, p. 157–188

Black, W. W., 1952, The origin of the supposed tufa bands in Carboniferous reef limestones: Geol. Mag., v. 89, p. 195–200

—— 1954, Diagnostic characters of the Lower Carboniferous knoll-reefs in the north of England: Leeds Geol. Assoc. Trans., v. 6, p. 262–297

Cayeux, L., 1935, Les roches sédimentaires de France: Roches carbonatées: Paris, Masson et Cie, p. 321–463

Cloud, P. E. Jr., 1952, Facies relationships of organic reefs: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 2125–2149

Cope, F. W., 1933, The Lower Carboniferous succession in the Wye Valley region of north Derbyshire: Manchester Geol. Assoc. Jour., v. 1, p. 125–145

—— 1939, The mid-Viséan (S2–D1) succession in north Derbyshire and north-west England: Yorkshire Geol. Soc. Proc., v. 23, p. 60–65

Cullis, O. G., 1904, The mineralogical changes observed in the cores of the Funafuti borings, sec. 14, p. 392–420 *in* Armstrong, H. E. *et al.*, The Atoll of Funafuti; Roy. Soc. London, Harrison and Sons, London, 428 p.

Dietz, R. S., and Menard, H. W., 1951, Origin of abrupt change in slope at continental shelf margin: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 1994–2016

Flower, R. H., 1942, Environment of early Paleozoic nautiloids: Nat. Research Council, Div. Geology and Geography Ann. Rept., v. 2, p. 37–41

Gardiner, J. S., 1931, Coral reefs and atolls: Macmillan, London, 181 p.

Hadding, A., 1941, The pre-Quaternary sedimentary rocks of Sweden, VI. Reef limestones: Lunds. Univers. Arrs. N.F., Avd. 2, Bd. 37, Nr. 10

—— 1950, Silurian reefs of Gotland: Jour. Geology, v. 58, p. 402–409

Henson, F. R. S., 1950, Cretaceous and Tertiary reef limestones and associated sediments in Middle East: Am. Assoc. Petroleum Geologists Bull., v. 34, p. 215–238

Hesse, R., Allee, W. C., and Schmidt, K. P., 1951, Ecological animal geography: John Wiley and Sons, Inc., New York

Hill, D., 1938, A monograph on the Lower Carboniferous rugose corals of Scotland: pt. 1, Palaeont. Soc., p. 1–78

Hudson, R. G. S., and Cotton, G., 1944, The Lower Carboniferous rocks in a boring at Alport, Derbyshire: Yorkshire Geol. Soc. Proc., v. 25, p. 254–330

Illing, L. V., 1954, Bahaman calcareous sands: Am.

Assoc. Petroleum Geologists Bull., v. 38, p. 1–95

Johnson, J. H., 1940, Algae and algal limestones from the Pennsylvanian of central Colorado: Geol. Soc. America Bull., v. 57, p. 571–596

—— 1946, Lime-secreting algae from the Pennsylvanian and Permian of Kansas: Geol. Soc. America Bull., v. 57, p. 1087–1120

Krumbein, W. C., and Garrels, R. M., 1952, Classification of chemical sediments: Jour. Geology, v. 60, p. 1–34

Kuenen, P. H., 1935, Geological interpretation of the bathymetrical results: The SNELLIUS Expedition, v. 5, Geol. Results, pt. 1, Leyden Brill, 123 p.

Newell, N. D., 1955a, Depositional fabric in Permian reef limestones: Jour. Geology, v. 63, p. 301–309

—— 1955b, Permian pelecypods of east Greenland: Meddelelser om Grønland, v. 110, p. 5–36

Newell, N. D., Rigby, J. K., Fischer, A. G., Whiteman, A. J., Hickox, J. E., and Bradley, J. S., 1953, The Permian reef complex of the Guadalupe Mountains region, Texas and New Mexico: W. H. Freeman and Co., San Francisco, 236 p.

Parkinson, D., 1943, The age of the reef limestones in the Lower Carboniferous of north Derbyshire: Geol. Mag., v. 80, p. 121–131

—— 1947, The Lower Carboniferous of the Castleton district: Yorkshire Geol. Soc. Proc., v. 27, p. 99–124

—— 1953, The Carboniferous limestones of Treak Cliff, Derbyshire with notes on the structure of the Castleton reef-belt: Geologists Assoc. London Proc., v. 64, p. 251–268

Pettijohn, F. J., 1949, Sedimentary rocks: New York, Harper and Sons, 526 p.

Plumley, W. J., and Graves, R. W., 1953, Virgilian reefs of the Sacramento Mountains, New Mexico: Jour. Geology, v. 61, p. 1–16

Revelle, R., 1934, Physico-chemical factors affecting the solubility of calcium carbonate in sea water: Jour. Sed. Petrology, v. 4, p. 103–110

Shirley, J., and Horsefield, E. L., 1940, The Carboniferous Limestone of the Castleton-Bradwell area, north Derbyshire: Geol. Soc. London Quart. Jour., v. 96, p. 271–299

Shrock, R. R., and Twenhofel, W. H., 1953, Principles of invertebrate paleontology: 2d ed., New York, McGraw-Hill Book Co., 816 p.

Smith, C. L., 1940, The Great Bahama Bank: Jour. Marine Research, v. 3, p. 147–189

Stehli, F. G., 1954, Lower Leonardian Brachiopoda of the Sierra Diablo: Am. Mus. Nat. History Bull., v. 105, p. 257–358

Sverdrup, H. U., Johnson, M. W., and Fleming, R. H., 1942, The oceans, their physics, chemistry, and general biology: New York, Prentice-Hall, Inc., 1060 p.

Vaughan, T. W., 1911, Physical conditions under which Paleozoic coral reefs were formed: Geol. Soc. America Bull., v. 22, p. 238–252

Walther, J., 1885, Die gesteinsbildenden Kalkalgen des Golfs von Neapel und die Entstehung structurloser Kalke: Zeitsch. Deutsch, Ges., v. 37, p. 329–357

Wilson, W. B., 1950, Reef definition: Am. Assoc. Petroleum Geologists Bull., v. 34, p. 181

Wood, A., 1941, "Algal dust" and the finer-grained varieties of Carboniferous limestone: Geol. Mag., v. 78, p. 192–200

—— 1942, The algal nature of the genus *Koninckopora* Lee; its occurrence in Canada and western Europe: Geol. Soc. London Quart. Jour, v. 98, p. 205–222

Zobell, C. E., 1946, Studies on redox potential of marine sediments: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 477–513

DEPARTMENT OF COLONIAL GEOLOGICAL SURVEY
KUCHING, SARAWAK

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, OCTOBER 24, 1956

TABLE 3.—LIST OF FOSSILS

Species	Shelf	Marginal Shelf and Back Reef		Reef	Fore Reef
		Castleton	Earl Sterndale		
CORALS					
Amplexus coralloides J. Sowerby	..	..	..	..	x
Caninia-benburbensis Lewis	x	x	x	..	x
C. sp.	x	x	x	..	x
Chaetetes depressus (Fleming)	..	..	..	x	..
C. septosus (Fleming)	x	x	x	..	x
Clisiophyllum cf. *keyserlingi* M'Coy	..	..	..	..	x
Dibunophyllum bipartitum (M'Coy)	x	x	x	x	x
D. bourtonense Garwood and Goodyear	x	x	x	x	x
Diphyphyllum sp.	..	..	..	..	x
Lithostrotion basaltiforme (Phillips)	x	x	x	..	..
L. irregulare (Phillips)	x	x	x	x	x
L. junceum (Fleming)	x	x	x	..	x
L. martini Edwards & Haime	x	..	..	..	..
L. m'coyanum Edwards & Haime	x	x	x	..	..
Michelinia sp.	..	..	..	x	..
Palaeosmilia murchisoni Edwards & Haime	x	x	x	x	x
P. regia (Phillips)	..	..	..	..	x
Syringopora sp.	x	x	x	..	x
BRACHIOPODS					
Actinoconchus planosulcatus (Phillips)	..	..	..	x	x
Athyris expansa (Phillips)	..	x	x	x	x
A. lamellosa (Leveillé)	..	..	..	..	x
A. sp.	x	x	x	..	..
Avonia youngiana (Davidson)	..	x	x	x	x
Brachythyris integricostata (Phillips)	..	..	..	..	x
B. ovalis (Phillips)	..	..	..	..	x
B. pinguis (J. Sowerby)	..	..	x	x	x
B. planicosta (M'Coy)	..	..	..	..	x
Buxtonia scabricula (J. Sowerby)	..	x	x	x	x
Camarotoechia pleurodon (Phillips)	..	..	..	..	x
Chonetes papilionacea Phillips	x	x	x	x	x
Cliothyridina roysii (Leveillé)	..	..	..	..	x
Davidsonina septosa (Phillips)	x	x	x	..	x
D. septosa var. *transversa* Jackson	..	..	..	..	x
Daviesiella aff. *comoides* (J. Sowerby)	x	x	x	..	..
Dictyoclostus antiquatus (J. Sowerby)	..	..	..	x	x
D. aff. *costatus* (J. Sowerby)	..	..	..	x	..
D. hindi (Muir Wood)	..	..	..	x	x
D. hindi var. *wettonensis* (Muir Wood)	..	..	..	..	x
D. insculptus (Muir Wood)	..	x	..	x	x
D. multispiniferus (Muir Wood)	..	..	..	..	x
D. pinguis (Muir Wood)	..	..	..	..	x
D. pugilis (Phillips)	..	x	..	x	x
D. sulcatus (Phillips)	..	..	..	x	x
D. sp.	..	x	x	..	..
Dielasma hastata (J. de C. Sowerby)	..	x	x	x	x
Echinoconchus elegans (M'Coy)	..	..	..	..	x
E. punctatus (J. Sowerby)	..	x	x	x	x
E. subelegans (Thomas)	..	..	..	..	x
E. venustus (Thomas)	..	..	..	..	x

TABLE 3.—*Continued*

Species	Shelf	Marginal Shelf and Back Reef		Reef	Fore Reef
		Castleton	Earl Sterndale		
Eomarginifera derbiensis (Muir Wood)	..	..	..	x	x
E. longispina (J. Sowerby)	..	..	..	..	x
E. minuta (Muir Wood)	..	..	..	..	x
E. pseudoplicatilis (Muir Wood)	..	..	..	x	..
E. setosus (Phillips)	..	..	..	x	x
E. triquetra (Muir Wood)	..	..	..	x	x
E. sp.	..	x	x	..	..
Gigantoproductus maximus (M'Coy)	x	x	x	..	..
G. spp.	..	x	x	x	x
Krotovea aculeata (J. Sowerby)	..	..	..	..	x
K. spinulosa (J. Sowerby)	..	..	x	x	x
Leptaena analoga Phillips	..	..	x	x	x
Linoproductus cora (d'Orbigny)	..	..	x	x	x
L. corrugatus (M'Coy)	..	..	..	..	x
L. hemisphericus (J. Sowerby)	x	x	x	x	x
L. undatus (Defrance)	..	..	x	x	x
Martinia glabra (J. Sowerby)	..	x	x	x	x
Overtonia fimbriata (J. de C. Sowerby)	..	x	x	x	x
Phricodothyris lineata (J. Sowerby)	..	..	x	x	x
Plicatifera aculeata (Martin)	..	..	..	x	..
P. carringtoniana (Davidson)	..	..	..	x	x
P. mesoloba (Phillips)	..	x	..	x	x
P. plicatilis (J. de C. Sowerby)	..	..	..	..	x
Proboscidella proboscideus (de Verneuil)	..	..	..	x	..
Productus concinnus (J. Sowerby)	..	..	..	x	x
P. productus var. *productus* Muir Wood	..	..	..	..	x
Pugnax accuminatus (J. Sowerby)	..	..	..	..	x
P. pugnus (Martin)	..	x	x	x	x
Pustula magnituberculata (M'Coy)	..	..	..	..	x
P. pustulosa (Phillips)	..	..	..	..	x
P. pyxidiformis (de Koninck)	..	..	..	..	x
Rhipidomella michelini (Leveillé)	..	..	..	x	x
Schellwienella crenistria (Phillips)	..	x	x	x	x
Schizophoria keyserlingiana (de Koninck)	..	..	..	..	x
S. resupinata (Martin)	..	x	x	x	x
Sinuatella sinuata (de Koninck)	..	x	..	..	x
Spirifer attenuatus J. Sowerby	..	..	..	..	x
S. bisulcatus J. Sowerby	..	x	x	x	x
S. convolutus Phillips	..	..	..	x	x
S. duplicicostus Phillips	..	..	..	x	..
S. duplicostatus M'Coy	..	..	..	..	x
S. striatus (Martin)	..	x	..	x	x
S. triangularis (Martin)	..	..	..	..	x
S. sp.	..	x	x	..	..
Spiriferellina cristata (J. Sowerby)	..	..	..	..	x
S. sp.	..	..	..	x	..
Striatifera striata (Fischer)	..	..	..	x	x
Syringothyris cuspidata (J. Sowerby)	..	..	..	..	x

TABLE 3.—*Continued*

Species	Shelf	Marginal Shelf and Back Reef		Reef	Fore Reef
		Castleton	Earl Sterndale		
PELECYPODS					
Allorisma sp.	..	..	..	..	x
Aviculopecten clathratus (M'Coy)	..	..	..	x	x
A. fimbriatus (Phillips)	..	..	..	x	..
A. murchisoni (M'Coy)	..	x	..	x	x
A. nobilis (de Koninck)	..	..	..	..	x
A. aff. *planoclathratus* (M'Coy)	..	..	..	x	..
A. plicatus (J. Sowerby)	..	..	..	..	x
A. semicostatus (Portlock)	..	..	..	..	x
A. tabulatus (M'Coy)	..	..	..	x	x
Cardiomorpha orbicularis (M'Coy)	..	..	..	..	x
Conocardium aleforme (J. Sowerby)	..	..	..	..	x
Crenipecten semicircularis (M'Coy)	..	..	..	..	x
Edmondia aff. *oblonga* (Portlock)	..	..	..	..	x
E. unioniformis (Phillips)	..	..	..	..	x
E. rudis M'Coy	..	..	..	x	..
E. sp.	..	..	x	..	..
Eumicrotis hemisphericus (Phillips)	..	..	..	..	x
Leiopteria grandis Hind	..	..	..	..	x
L. sp.	..	x	..	x	..
Lithodomus lingualis (Phillips)	..	..	..	x	x
Modiola aff. *patula* M'Coy	..	..	..	..	x
Myalina peralata de Koninck	..	..	..	..	x
M. pernoides (Portlock)	..	..	..	..	x
Mytilomorpha rhombea (Phillips)	..	..	..	..	x
Palaeolima sp.	..	..	..	..	x
Parallelodon bistriatus (Portlock)	..	..	x	..	x
P. cingulatus M'Coy	..	..	..	..	x
P. cf. *obtusus* (Phillips)	..	..	..	..	x
P. ornatissimus de Koninck	..	..	..	..	x
P. squamifer (Phillips)	..	..	..	..	x
P. sp.	..	x	..	..	..
Pinna sp.	..	..	..	..	x
Posidoniella gibbosa Hind	..	..	..	x	x
P. vetusta (J. Sowerby)	..	..	..	x	x
Protoschizodus cf. *fragilis* (M'Coy)	..	..	..	..	x
P. trigonalis de Koninck	..	..	..	..	x
Pseudamusium concentrico-lineatum Hind	..	x	..	..	..
Pterinopecten dumontianus (de Koninck)	..	..	..	x	..
P. granosus (J. Sowerby)	..	..	..	x	x
P. tesselatus (Phillips)	..	..	..	x	x
Sanguinolites angulatus de Koninck	..	..	..	..	x
S. aff. *tricostatus* (Portlock)	..	..	..	..	x
S. sp.	..	x	..	..	..
Spathella aff. *cylindrica* (M'Coy)	..	..	..	..	x

TABLE 3.—*Continued*

Species	Shelf	Marginal Shelf and Back Reef		Reef	Fore Reef
		Castleton	Earl Sterndale		
GASTROPODS					
Aclisina sp.	..	..	..	..	x
Bellerophon hiulcus (Martin)	..	..	..	..	x
B. cf. *munsteri* d'Orbigny	..	..	..	..	x
B. tenuifascia J. Sowerby	..	..	..	..	x
B. sp.	x	x	x	x	x
Capulus sp.	..	..	..	..	x
Euomphalus crotalostomus M'Coy	..	..	..	..	x
E. pentangulatus J. Sowerby	..	x	x	..	..
Euphemus sp.	..	..	..	..	x
Flemingia carbonaria Meek and Worthen	..	..	..	..	x
F. aff. *hiseringiana* de Koninck	..	..	..	..	x
Lepetopsis cf. *phillipsi* de Koninck	..	..	..	..	x
L. retrorsus (Phillips)	..	..	..	..	x
Loxonema sp.	..	..	..	..	x
Macrochelina cf. *tumida* de Koninck	..	..	..	..	x
Mourlonia naticoides (de Koninck)	..	..	..	..	x
M. sp.	..	..	..	x	..
Murchisonia cf. *humboldtiana* de Koninck	..	..	..	..	x
Naticopsis ampliata (Phillips)	..	..	..	..	x
N. aff. *consimilis* de Koninck	..	..	..	..	x
N. globosa (Hoeninghaus)	..	..	..	..	x
N. sp.	..	x	x	x	..
Phymatifer pugilis (Phillips)	..	..	..	..	x
Platyschisma helicoides (J. Sowerby)	..	..	..	..	x
Ptychomphalus conimorphus de Koninck	..	..	..	..	x
Straparollus aequalis (J. Sowerby)	..	..	..	..	x
S. cf. *fallax* de Koninck	..	..	..	..	x
S. heliciformis de Koninck	..	..	..	..	x
S. ineptus de Koninck	..	..	..	..	x
S. laevigatus (Leveillé)	..	..	..	..	x
S. aff. *placidus* de Koninck	..	..	..	..	x
S. sp.	..	x	x	..	..
Turbonellina pulchra (de Koninck)	..	..	..	..	x
Turbonitella biserialis (Phillips)	..	..	..	..	x
NAUTILOIDS					
Aclinoceras sp.	..	..	..	..	x
Coelogasteroceras sp.	..	..	..	..	x
Coelonautilus planotergatus Hyatt	..	..	..	..	x
Orthoceras wrighti Haughton	..	..	..	..	x
O. sp.	..	..	..	x	x
Rayonoceras sp.	..	..	..	..	x
Stroboceras sulcatum (J. Sowerby)	..	..	..	..	x
Vestinautilus crateriformis Foord	..	..	..	..	x
Indet. nautiloid	..	x	..	..	..

TABLE 3.—*Continued*

Species	Shelf	Marginal Shelf and Back Reef		Reef	Fore Reef
		Castleton	Earl Sterndale		
GONIATITES					
Beyrichoceras castletonense (Bisat)	..	..	..	..	x
B. cf. *delicatum* Bisat	..	..	..	..	x
B. micronotum (Phillips)	..	..	..	..	x
B. aff. *phillipsi* Bisat	..	..	..	..	x
B. cf. *rectangularis* Bisat	..	..	..	..	x
B. cf. *submicronotum* Bisat	..	..	..	..	x
B. sp.	..	..	..	x	..
Dimorphoceras sp.	..	..	..	..	x
Goniatites cf. *antiquatus* Bisat	..	..	x	..	x
G. hudsoni Bisat	..	..	..	..	x
G. maximus Bisat	..	..	..	..	x
G. maximus var. *b* Bisat	..	..	..	..	x
G. sp.	..	x	x	..	..
Nomismoceras cf. *vittigerum* (Phillips)	..	..	..	..	x
SPONGES					
Microspongia sp. nov.	..	..	..	x	..
Lithistid sponge gen. et sp. nov.	..	..	..	x	..
TRILOBITES					
Brachymetopus ouralicus (de Verneuil)	..	..	..	..	x
Griffithides longiceps Portlock	..	..	..	x	..
G. seminiferus (Phillips)	..	..	..	..	x
Phillipsia derbiensis (Martin)	..	..	..	..	x
P. eichwaldi (Fischer)	..	..	..	..	x
P. cf. *gemmulifera* (Phillips)	..	..	..	..	x
P. sp.	..	x	x	..	..
Indet. trilobites	..	x	x	x	x
OSTRACODS	x	x	x	x	x
BRYOZOANS					
Fenestella sp.	..	x	x	x	x
Fistulipora minor M'Coy	..	..	x	x	x
Penniretepora pluma (Phillips)	..	..	..	x	x
CRINOIDS	x	x	x	x	x
BLASTOIDS	..	..	..	..	x
FORAMINIFERS	x	x	x	x	x
ALGAE					
Calcisphaera sp.	x	x	x	x	x
Girvanella ducii Wethered	..	x	x	x	x
G. incrustans Wethered	..	x	x	x	x
G. nicholsoni (Wethered)	..	x	..	x	..
G. staminea Garwood	..	x	x	x	..
G. sp.	x	x	..	x	x
Koninckopora inflata (de Koninck)	x	x	x	..	x
Ortonella kershopensis Garwood	..	..	..	x	..
O. tenuissima Garwood	..	x	..	..	..
Solenopora sp.	..	..	x	..	..

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SEDIMENTARY BRECCIAS IN THE SOUTHEAST MISSOURI LEAD DISTRICT

BY F. G. SNYDER AND J. W. ODELL

ABSTRACT

Sedimentary breccias are important ore-controlling structures in the lower half of the Bonneterre Formation in the southeast Missouri lead district. Four major breccia zones, each embracing over 60,000,000 tons of rock, have been delineated. Individual ore bodies in the breccia zones range from over 6000 feet long and containing several million tons to a few hundred feet with several thousand tons.

The breccias formed along the flanks of calcarenite ridges by submarine slides into near-by depositional basins. Extensive mine workings and closely spaced diamond drilling in the breccia bodies permit reconstruction of entire slides and determination of facies relationships leading to their development.

Lithofacies studies indicate that the four basic types of material composing the lower part of the Bonneterre Formation—argillaceous mud, sand-size sediment, recognizable algal deposits, and carbonate mud—may be referred to environments of deposition which are dominated by two elements, depositional ridges and basins. The ridges are composed for the most part of calcarenite and reef deposits, the basin sediments of fine-grained carbonate and argillaceous muds. Differential compaction led to oversteepening of slopes in the gradational zone between the ridge and basin deposits and initiated movement of unconsolidated and partially lithified sediments. Repeated slides at fairly close stratigraphic intervals account for great thicknesses of breccia in the lower part of the formation.

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Introduction

General Statement

Breccias occur in sedimentary rocks of all ages. They include great masses attributed to submarine slump and gliding and smaller bodies ascribed to intrastratal flow as well as normal depositional types. Published descriptions of many of the breccia units and their mode of formation are based almost entirely on study of surface exposures.

Extensive zones of sedimentary breccia in the southeast Missouri lead district present a unique opportunity for three-dimensional study of the breccias, their relationships to other types of lithology, the environments in which they formed, and their mode of formation. Within an area of approximately 60 square miles several square miles of open stopes and approximately 60,000 cored, diamond drill holes make unusually complete information available.

During the past decade, geologists of the St. Joseph Lead Company, under Dr. John S. Brown, have been making intensive studies of the disseminated lead deposits of southeast Missouri. The major geologic features of the district were outlined in a progress report by Ohle and Brown (1954), and several phases of applied geology were reported by Snyder and Emery (1956).

This paper presents further studies of the lithology of the ore-bearing Bonneterre Dolomite and recognizes primary sedimentary breccias as important units in the formation and as major ore-controlling structures.

Acknowledgments

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Geologic Setting

General geology.—Upper Cambrian formations overlying the maturely dissected Precambrian surface include, from the base upward, the Lamotte Sandstone, Bonneterre Dolomite, Davis Shale, and Derby-Doerun Dolomite. The Lamotte Sandstone has a maximum thickness of 425 feet within the mining district and feathers out on the flanks of the higher Precambrian knobs.

The Bonneterre Formation averages about 400 feet thick. Lower units may be absent where it caps Precambrian igneous knobs. The formation is composed predominantly of dolomite within the district, with some limestone "islands" between, above, or below mineralized blocks; outside the productive area large parts of the formation are limestone. Most of the ore has come from the lower 250 feet of the formation; the upper units have produced significant amounts of ore only in restricted areas.

The central productive area of the lead district is bounded by the Simms Mountain fault on the southwest side, the Big River fault on the northwest, and the Farmington anticline on the northeast (Wagner, 1947). A downdrop of approximately 500 feet on the north side of the Simms Mountain fault has superimposed a southwest regional dip of nearly 2° on the district.

Within the mined area a complex fracture set with a predominantly northwest trend is

expressed in a number of faults with 40–100 feet displacement and many lesser breaks. Many faults show evidence of both pre-ore and post-ore movement. Trend of ore bodies is controlled for the most part by primary sedimentary structures, but at intersections of faults and sedimentary structures, ore bodies may be enlarged.

The ore mineralogy is relatively simple (Tarr, 1936). Galena is the main mineral. Sphalerite is present in places and locally may exceed galena. Minor amounts of chalcopyrite and siegenite are recognized. Marcasite, pyrite, and calcite are common. Secondary lead and copper minerals occur in minor amounts.

Subdivision of the Bonneterre Formation.—The Bonneterre Formation was divided into lithic-stratigraphic units by J. E. Jewell and R. E. Wagner prior to 1947. Their classification, with minor changes, provides the basis for the present core-logging system and for underground mapping. The formation is divided into zones numbered 1 through 19, from the top down. The major units are shown in Figure 1. Detailed descriptions of units are given by Ohle and Brown (1954, p. 205).

All zones are generalized lithologic types. The zones are predominantly tan crystalline calcarenite[1] alternating with gray or brown shaly carbonate of varied texture. The lower units, 10 through 19, vary considerably in thickness or may be absent. Where a unit thins, the adjacent one thickens, so that the over-all thickness of the formation is about the same throughout the central productive area.

The 19 bed is a sandy dolomite regarded as the transitional phase between the Lamotte quartz sand and the Bonneterre carbonates. The transitional unit is a member of the Bonneterre Formation and in many areas is separated by a sharp hiatus from the underlying sandstone. Quartz sand of the transition unit represents reworked Lamotte. Quartz-bearing beds of the 19 unit intertongue with quartz-free Bonneterre units on the flanks of depositional ridges.

The 15, 10, and 5 units are composed predominantly of tan, crystalline calcarenite; in many places the 10 and 5 are oölitic. The 15 and 10 units are markedly lenticular, strongly cross-bedded, and variable in thickness. The 12 and 7 beds are gray to brown shaly dolomite of varied lithologies. The 7 bed represents the main period of algal reef formation (Ohle and Brown, 1954). Where algal formation is restricted or lacking, the unit is composed of gray shaly dolomite like the 12 unit.

Calcitic limestone, designated as the 11 bed, rarely is cut by mine workings. Away from mine workings and off the ore-bearing structures it is penetrated by many drill holes, most commonly within the upper part of the 12 and the lower part of the 7, although in places the entire section from the top of the 19 to the base of the 5 is limestone. Lithologic character of the limestone has had little detailed study, but most of the textural variations occurring in dolomite also are found in limestone.

One unusual rock type, classed as the 9 unit, is a sedimentary breccia of limited areal and stratigraphic extent. It represents a primary or penecontemporaneous product of slump or intrastratal flow in shaly 12 or 7 beds. It is most common in the 12 but may occur at any position in the lower 200 feet of the formation, except in the tan crystalline units.

Lithology of the Lower Part of the Bonneterre Formation

Classification of Sediments

The tetrahedral classification.—The Lamotte Sandstone and the Bonneterre Dolomite represent a normal orthoquartzite-limestone sequence (Pettijohn, 1949, p. 453). This paper is concerned with the carbonate member of the sequence, particularly with the lower 250 feet of the Bonneterre.

All textural and compositional variations in the lower 250 feet of the Bonneterre Formation within the central productive area can be described in terms of a tetrahedron, showing all possible rock types and gradations (Fig. 2).

The four types of material which form the end members in the calssification are:

(1) Argillaceous muds composed for the most part of illite. They may be either black or green, depending upon environment of deposition and diagenetic history.

(2) Sand-size and coarser sediments composed of carbonate with or without detrital rock and quartz grains. The carbonate includes oölites, pisolites, organic debris, and coherent,

[1] There is little general agreement on the terminology of carbonate rocks. The term limestone is used in this paper for rocks composed chiefly of calcite; the term dolomite for those composed chiefly of high magnesian carbonate. Grade limits of sediment are expressed by the terms calcarenite, and calcilutite, but throughout the paper these terms are used only to designate dolomitic rocks, specifically, dolomitized calcarenites and dolomitized calcilutites.

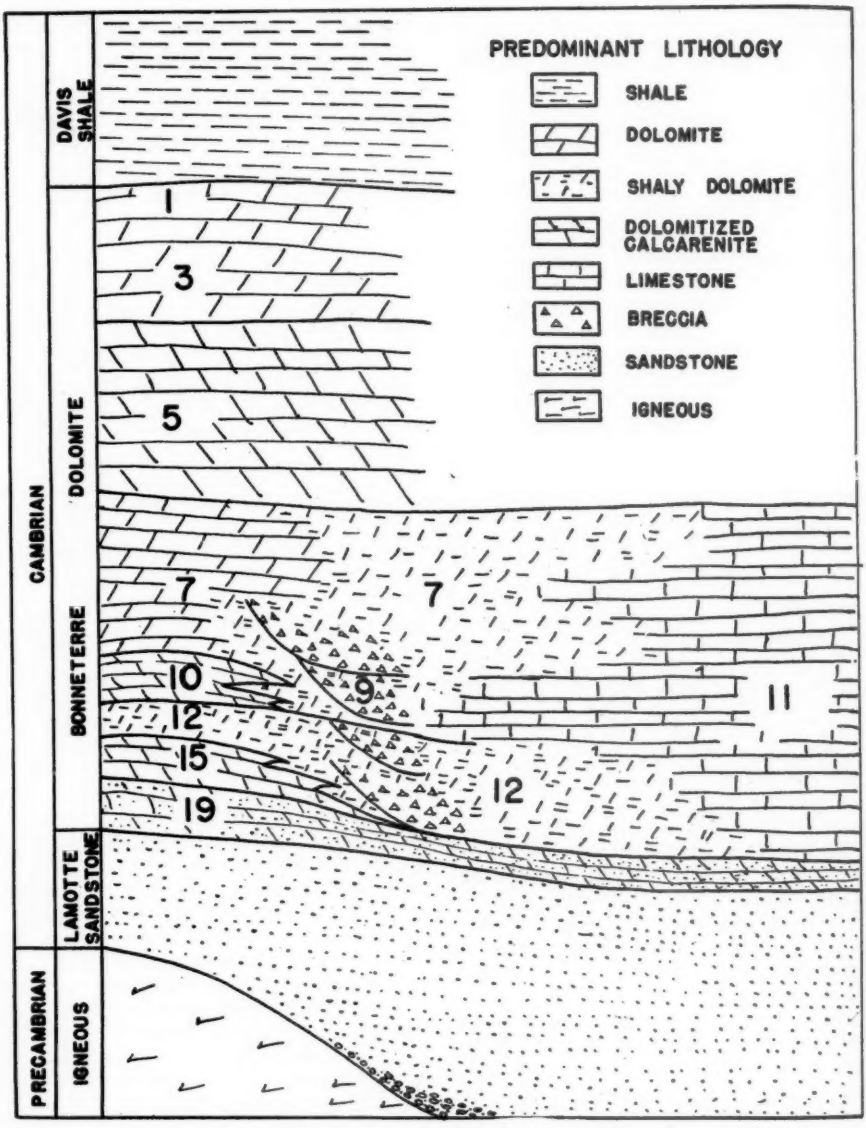


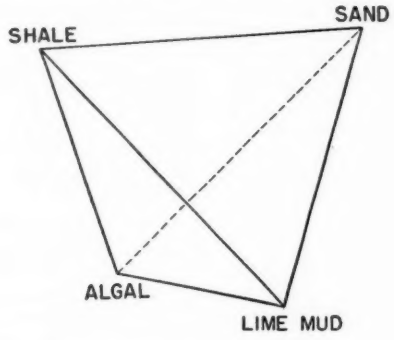
FIGURE 1.—GENERALIZED STRATIGRAPHIC SECTION

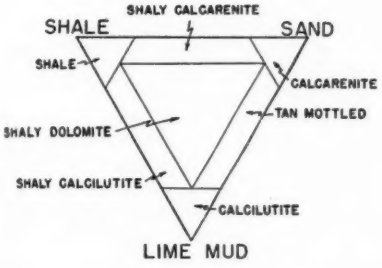
slightly indurated aggregates of reworked carbonate.

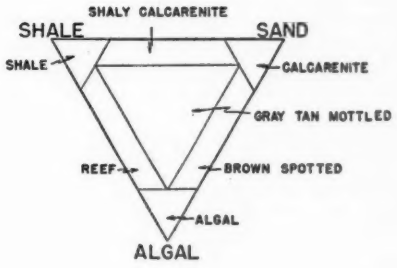
(3) Carbonate muds of uncertain origin. They may include fine-grained biochemical and chemical precipitates as well as detrital lime mud.

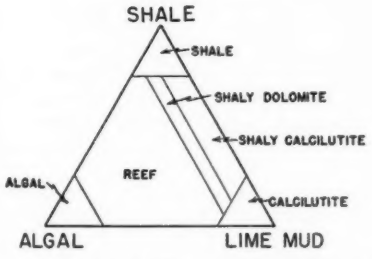
(4) Organic deposits composed mainly of algal structures, although other unidentifiable fossil forms probably are present.

Diagenetic changes.—The cataloging of carbonate rock types in terms of proportions of sediment end members may be hampered by the effects of dolomitization and recrystallization, but such effects do not preclude develop-









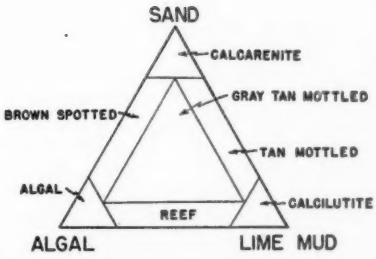


FIGURE 2.—TETRAHEDRAL CLASSIFICATION OF LOWER BONNETERRE SEDIMENTS

ment of a reliable and useful classification. Although diagenetic processes may have destroyed skeletal remains and modified grain size so that details of minor structures and textures are lacking or are obscure in thin section or in limited exposure, gross sedimentary features of the Bonneterre Formation are sufficiently well preserved to permit reconstruction of depositional environment and interpretation of facies relationships.

The prime source of the carbonate is unimportant to the classification used. Formation of the carbonate grains may be due to chemical precipitation, organic activity, or aggregation (Illing, 1954, p. 1–95). The grains, here regarded as allochthonous carbonate, may be transported and deposited as sand, silt, or mud. These may be well sorted, partially sorted, or unsorted; the degree of sorting is controlled by the transporting agent and the depositional environment. The well-sorted sediments form deposits of uniform grain size; the poorly sorted and unsorted sediments form rocks of mixed lithology and varied texture. Diagenetic changes modify the original sediment, but, because dolomitization and recrystallization are

selective processes, significant features, indicative of the character of the original material, are preserved in any suite of carbonate rocks.

Thin-section study alone cannot define the character of the original sediment. In the Bonneterre Formation a calcarenite may appear microscopically as a mosaic of equigranular anhedral dolomite crystals. Dolomite crystals comprising oölites, fossil relicts, carbonate aggregates, or cementing material may be identical in size and appearance (Pl. 1, fig. 2). They differ in size from the original grains. Grain boundaries of oölites and carbonate aggregates vary in degree of preservation. Original particles are difficult to recognize in thin section; they are much more apparent in hand specimen and drill core (Pl. 1, figs. 1 and 6; Pl. 2, fig. 3). Original grain size is indicated by ghosts of oölites and carbonate aggregates, by quartz grains, and by clay pellets. Gross features of exposures such as cross-bedding, scour and fill, and marked lenticularity of beds indicate a clastic origin. Study of facies relationships provides knowledge of the depositional environment, giving an additional check on interpretation of effects of diagenesis.

Organic structures are apparent megascopically, but recrystallization has destroyed all internal structure. In thin section, dolomite within organic forms is indistinguishable from adjacent interstitial material (Pl. 2, fig. 2); in hand specimen and mine exposures gross organic structures are striking in appearance and easily recognized. The destruction of the internal organic structure is accompanied by the formation of patches of fine-grained dolomite, which are indistinguishable microscopically from bedded calcilutite but differ markedly from it in gross relationships of bedding, associated lithologies, and depositional environment. This selective recrystallization imparts a mottled appearance to rocks of originally heterogeneous texture. The mottling is ubiquitous throughout the lower part of the Bonneterre Formation in beds of fossil debris, *in situ* organic reefs, and beds of unsorted clastic carbonate.

The common occurrence of mottling in particular types of sediment from certain depositional environments, and its absence in other types of sediment and depositional environments, is regarded as indicative of the character of the original sediment. Rocks which are uniform show well-preserved bedding, and distinctive textures or structures conform to bedded units. The rock consists of one end-member type or is a simple mixture consistent with a particular depositional environment. The rocks of variegated texture are massive, structureless units, or show vague or irregular bedding, and are a complex mixture of several end-member types. Within the Bonneterre Formation, rocks which are uniform in character following dolomitization and recrystallization were essentially uniform in character of sediment; rocks which have a mottled and variegated texture were formed from sediment which varied in grain size or type of material.

The common lithologies.—End-member types of sediment occur in relatively pure form in restricted environments. The shale constituent is present as green shale and mudstone in local sedimentary basins and as black, finely laminated units (Pl. 1, fig. 5). Both the green and black shale are composed of illite (Beaumont, 1953, unpub. Ph.D. thesis, Columbia Univ.). The black shale derives its color from finely divided pyrite; it also contains minor detrital quartz and authigenic feldspar. Most of the black shale units are less than a few inches thick and are associated with beds of organic detritus.

Various types of sand-size material make up large parts of the calcarenite units shown on Figure 1. Exposures of the calcarenites in mine workings show numerous scoured surfaces, strong cross-bedding, and marked lenticularity of beds, features indicative of deposition in shallow agitated waters. Individual grains may be oölites, pisolites, aggregates of carbonate, fossil relicts, quartz sand, and porphyry pebbles. Much of the carbonate sand probably represents abraded and recrystallized organic detritus, in which evidence of organic origin has been destroyed (Pl. 2, fig. 3). In some beds outlines of original grains have been obliterated completely; recognition of nature of the sediment is based on facies relationships and primary structural features.

The calcarenite units grade from clean, well-sorted sands to impure varieties, owing to admixture of lime silt and lime mud. Carbonate sand may be present within the reef-framework filling channels and interstices between individual algal structures. The calcarenites (and calcisiltites) may be interbedded with shales or calcilutites. The thin-bedded carbonate units alternating with very thin shale units commonly have a wavy bedded or nodular character.

Deposits composed of recognizable organic material alone are rare. Typically, the algal reefs of the 7 zone provided a rigid framework in which other sediments were entrapped. In the large reef masses only the genus *Collenia* has been identified (Ohle and Brown, 1954).

Material referred to the genus *Girvanella* is common in some beds, but this was transported and deposited as a clastic. Other organisms undoubtedly were present, but owing to dolomitization and recrystallization are unidentifiable. As stated earlier, internal structures of organisms have been destroyed, and a large percentage of the reef rock, of both organic and clastic origin, consists of microcrystalline calcilutite. Probably much of the calcilutite is a result of diagenetic processes but part may be due to original lime mud[2].

Beds of organic detritus, probably representing a coquina, and small patch reefs are common in the lower part of the formation. In many places these units lie disconformably on the underlying beds; they commonly occur with black shale; and they appear to be limited to a particular local sedimentary environment.

Calcilutite is an important constituent of depositional basins and also occurs as thin disjointed units interbedded with calcarenites and calcisiltites. In thin section the calcilutite is uniform, microcrystalline mosaic of anhedral dolomite. In hand specimen it is a dense aphanitic rock, locally called slime because of its apparent lack of crystallinity. Diagenetic effects may have modified grain size of the original sediment, but the identity and continuity of units has been preserved. There is no tendency for destruction or intermixing of thin alternating calcarenite and calcilutite beds by dolomitization and recrystallization. The uniform character throughout persistent thin beds, the occurrence in sedimentary basins, and the association with argillaceous muds leave no alternative to classifying this type of calcilutite as a carbonate mud. This conclusion is substantiated by the occurrence of occasional thin beds of quartz-free microcrystalline carbonate in the quartz-bearing calcarenites of the 19 zone.

The shale imparts a gray color to the carbonate rock. The coloring is due to fine-grained iron sulfide associated with the shale, not to carbonaceous material. As the proportion of shale increases, the tan rocks pass into gray dolomites. The gray dolomites consist of mixtures of two or more end-member types, and large parts of the 7 and 12 zones consist of mixtures of all four end members. The constituents—sand, argillaceous and carbonate mud, and organic material—may vary in any proportions. Consequently, the gray dolomites show an almost unlimited range of textural types due to diagenetic changes superimposed on sediments of mixed composition.

Most of the gray dolomites lie within the tetrahedron of the present classification in terms of both present character and original sediment. In many of the mixed rock types it is impossible to assess the relative importance of original lime mud relative to fine-grained products of recrystallization. Probably both are important constituents. Significant quantities of argillaceous mud suggest that lime mud was also present. Facies relationships and gross structural features, which indicate the sediments were deposited below wave base without being reworked by waves and currents, support the probability that lime mud was quantitatively important. Conversely, partially destroyed but recognizable grains now composed of calcilutite and the absence of bedding or other primary structural features in some calcilutite indicate that much of the microcrystalline carbonate is a product of diagenesis. The net effect of diagenesis on mixed sediments composed of several end-member types would be a shift in position within the tetrahedron toward the lime mud end member.

Environmental Control of Lithology

The Bonneterre sediments are deposits of a comparatively stable shelf zone. The Precambrian surface, except for the higher knobs, was buried beneath a blanket of Lamotte sand, which formed an undulating surface on which carbonate deposition began. The higher knobs were important in channeling currents which built sand and sandy carbonate bars and spits, which further emphasized the bottom irregularities.

Over this uneven bottom carbonate deposition took place in a variety of environments. The highs on the sea floor were periodically within the zone of water agitation, during which time sediments were reworked, the fine carbonate mud and silt were swept out into basin areas, and only clean, well-sorted carbonate sands remained over the highs. Size of particles may decrease along the strike where bars encroach into the basin. On the flanks of minor bars and banks, changing depth conditions resulted in a variety of overlap structures

[2] Newell, *et al.* (1953, p. xiv) expressed doubt concerning the quantitative importance of lime mud in reefs. The algal reefs of the Bonneterre Formation contain appreciable quantities of argillaceous mud, indicating that sheltered areas within the reef could receive fine-grained sediments. Lime mud of detrital or biochemical origin could have been and probably was deposited with the argillaceous mud.

and rapid change of facies. No evidence has yet been recognized that would indicate the bars and banks stood above water level. The highs on the sea bottom were the main sites of deposition of the carbonate sand and organic end-member types; the basin areas were deposition sites of the argillaceous mud and carbonate mud types. Intermittent fluctuation of water level or variation in wave activity during early Bonneterre time gave rise to an alternation of tan and gray carbonate rocks over the depositional ridges.

The ridges which began in late Lamotte and early Bonneterre time persisted to mid-Bonneterre time as topographic highs, over which the character of sediments differed markedly from the intervening basins. Trend and position of ridge axes shifted slightly with succeeding beds, but an approximate vertical continuity of the depositional environment was maintained in many areas throughout the first half of Bonneterre time (Ohle and Brown, 1954).

Local sedimentary basins within the district presented an entirely different set of conditions as compared with the highs. The basins received fine-grained lime mud and silt and argillaceous mud throughout most of early Bonneterre time. Rarely, if at all, did sediments of the basins rise above wave base. Times of exceptional wave or current activity periodically may have brought coarser sediments into the basins, but these were never reworked, sorted, or aerated as were sediments over the highs.

Along the margins of the sedimentary basins submarine slides during early Bonneterre time gave rise to extensive zones of carbonate and shale breccia. The breccia masses form a locally important zone of the formation; they are useful in paleogeographic interpretation; and they are host rocks for major lead-zinc ore bodies.

The Submarine Slides

Character of the Breccia

Lithology.—The breccias are composed of calcilutite fragments set in a fine- to medium-grained carbonate matrix (Pl. 1, fig. 3), or of green shale fragments with similar carbonate matrix.

Degree of brecciation may vary. Locally within the disturbed mass, beds may be crumpled and folded without fracturing, and considerable movement may be revealed by slipping along bedding planes and flow of material (Pl. 3, fig. 3). Where the rock is intensely broken, fragments vary from a fraction of an inch to several feet (Pl. 3, fig. 5, 6). Many calcilutite fragments represent disjointed segments of a given bed and have the thickness of the sedimentary unit from which they were broken. Most blocks are undeformed, showing no bending nor crumpling. Corners and edges of fragments are angular; the blocks have undergone little abrasion.

The matrix of the dolomite breccia consists of

Plate 1.—Details of Lithology

Figure 1.—Sawed diamond-drill core of dolomitized oölite. Light colored matrix is calcarenite. Width of core is 1½ inches.

Figure 2.—Photomicrograph of relict clastic particle. Texture of recrystallized particle and parts of matrix are identical. Cross nicols, magnification 60×.

Figure 3.—Sawed diamond-drill core of calcilutite breccia (9 bed).

Figure 4.—Photomicrograph of shaly calcilutite. Thin black shale seam and disseminated black shale in fine-grained carbonate. 60×.

Figure 5.—Sawed diamond drill core drilled parallel to bedding. Tan crystalline dolomite (calcarenite) above, black shale with detrital quartz grains, and organic detritus below.

Figure 6.—"Gray spotted" dolomite consisting of relict fossils and clastic carbonate with calcilutite matrix. Sawed diamond-drill core.

Plate 2.—Lithologic and Structural Features

Figure 1.—Sawed diamond-drill core of lower organic zone within 12 bed. Consists of organic material and calcilutite.

Figure 2.—Thin section of organic rock of 7 bed. Dark material on right is algal structure; light material on left is clastic carbonate. 60×.

Figure 3.—Thin section of "spotted" dolomite. Clusters of dark material represent "spots"; white patches are calcilutite. 60X. Compare with Plate 1, figure 6.

Figure 4.—Dip of beds on flank of calcarenite ridge. Chalk line *H-H* marks horizontal.

Figure 5.—Upper contact of slide and overlying beds, Schultz area. Top of slide marked by broken line. Galena and marcasite along contact.

Figure 6.—Base of slide, Schultz area. Same slide as shown in Plate 2, figure 5.

FIGURE 1

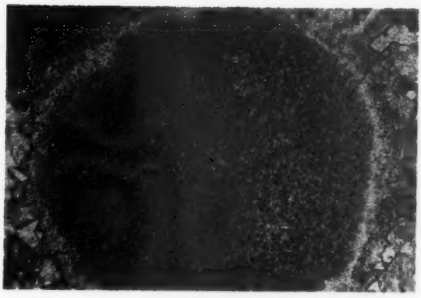


FIGURE 2

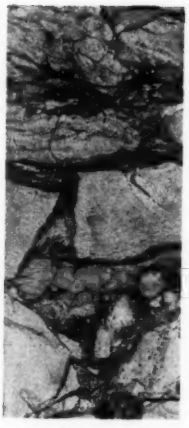


FIGURE 3

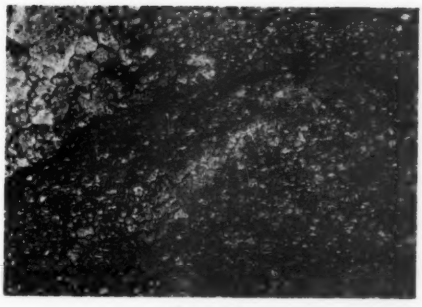


FIGURE 4

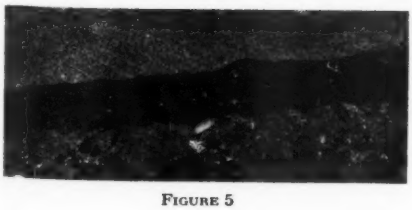


FIGURE 5

FIGURE 6

DETAILS OF LITHOLOGY

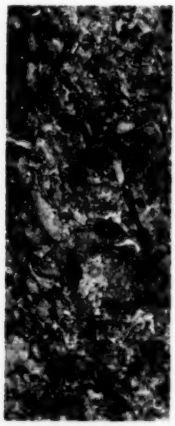


FIGURE 1

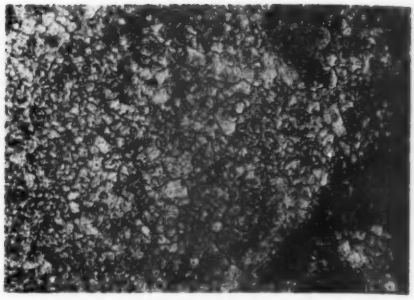


FIGURE 2

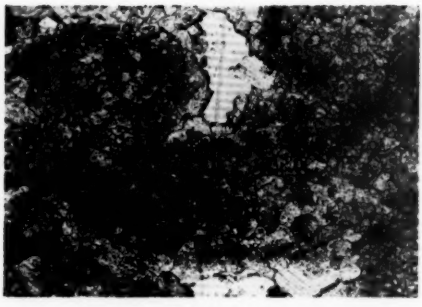


FIGURE 3

FIGURE 4

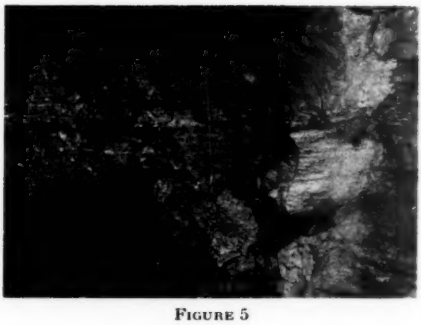


FIGURE 5

FIGURE 6

LITHOLOGIC AND STRUCTURAL FEATURES

FIGURE 1

FIGURE 2

FIGURE 3

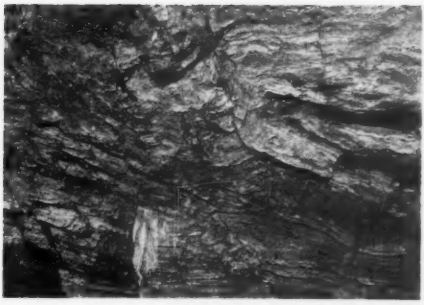


FIGURE 4

FIGURE 5

FIGURE 6

DETAILS OF BRECCIA MASSES

FIGURE 1

FIGURE 2

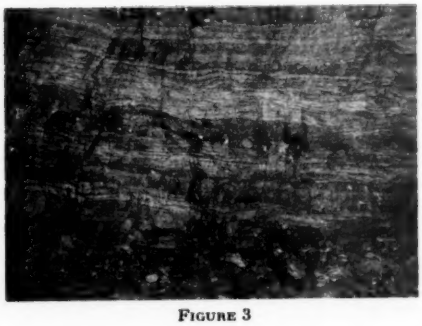


FIGURE 3

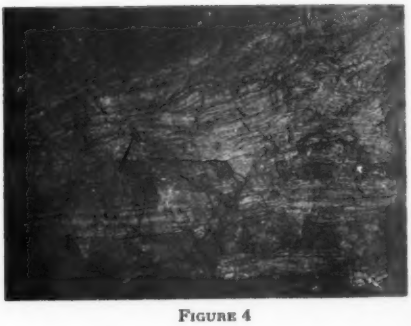


FIGURE 4

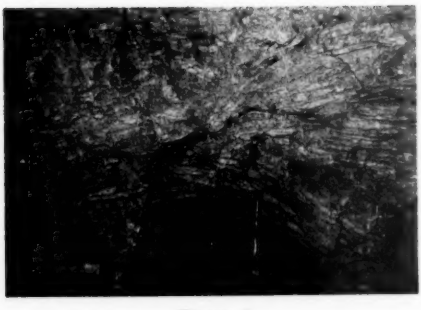


FIGURE 5

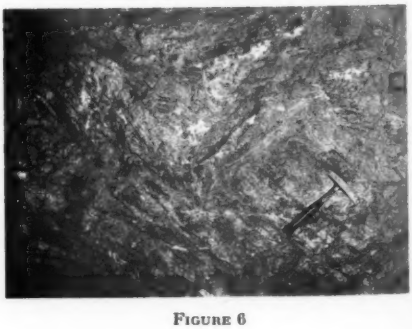


FIGURE 6

STAGES OF BRECCIATION

fine- to medium-grained crystalline brown dolomite with varying proportions of black shale. It includes carbonate of recognizable organic origin, calcarenite, calcisiltite, and black shale. Whereas the breccia fragments along the edges of the mass can be traced to unbroken beds of calcilutite outside the brecciated zone, the matrix material has lost its stratification and can be traced into units of undeformed rock only by its position relative to the more competent calcilutite beds (Pl. 3, fig. 5). Within the brecciated zone the matrix material has flowed around the breccia blocks, completely enclosing them. The calcilutite fragments float in the coarser-grained matrix.

The shale breccias are composed of irregular fragments of green shale set in a fine-grained carbonate matrix. Black shale, always present in some amount in the dolomite breccia, is absent in the shale breccia. The parent rock of the shale breccia was a green dolomitic shale with interbedded thin units of coarser nonshaly dolomite. The dolomitic shale contains large amounts, up to 50 per cent, of fine-grained carbonate evenly and intimately dispersed throughout the argillaceous bed. The coarser carbonate comprising the matrix is minor and may be so rare that the rock appears to be almost entirely shale, the brecciated character being apparent only when attitudes of laminae in breccia fragments are examined. Where shale units are thin, fragments may be folded; where they are thick, the ends of fragments may be blunted and distorted at contacts with adjacent fragments.

Both the dolomite breccia and the shale breccia are dense and tight; cavities or open spaces are rare. In the dolomite breccia, open spaces developed during brecciation have been filled by flow of the incompetent coarser-grained carbonate; in the shale breccia they have been filled by deformation of the shale fragments and to a minor extent by flow of carbonate.

Stratigraphic distribution.—Four large slide zones and many smaller ones are shown on Figure 3. Incomplete studies indicate that several other large slides are present adjacent to mined areas. Approximate size of the larger masses, as delineated from underground exposures and closely spaced drill holes, is shown in Table 1.

Within the disturbed areas, breccia is extensive, but a slide zone is not a single chaotic mass. Breccia at several different stratigraphic levels may be separated by undisturbed beds, and within a given stratigraphic zone, breccia bodies may be separated by unbroken beds. The areas outlined (Fig. 3) represent general zones of instability in which brecciation involving relatively small individual areas occurred repeatedly during early Bonneterre time.

Individual breccia bodies may be several thousand feet long, over a thousand feet wide, and up to 80 feet thick. A slide originating in higher beds may truncate one which had formed earlier in lower beds, resulting in continuous thicknesses of breccia up to 130 feet.

The slides are most common in the 12 zone but may occur at any stratigraphic position

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PLATE 3.—DETAILS OF BRECCIA MASSES

FIGURE 1.—Low-angle glide plane, Owl Creek slide. Chalk line on face shows trace of glide plane.

FIGURE 2.—Slickensides on base of glide plane as exposed on mine back. Slide descending from upper left diagonally to lower right. Beds on lower right below inked line are above glide plane. Width of exposure approximately 15 feet.

FIGURE 3.—Flowage of beds within slide mass. Deformation decreases downward due to frictional drag and greater compaction of sediments.

FIGURE 4.—Drag fold at toe of slide.

FIGURE 5.—Toe of slide where breccia passes into undeformed basin sediments. Lower units marked by chalk. Bed marked by pick folded but not broken. Note underthrusting in broken calcilutite bed (cross-hatched).

FIGURE 6.—Intense brecciation in lower part of Leadwood slide. Note alignment of breccia blocks marking completely healed thrust zone (diagonally up to right, center of photograph).

PLATE 4.—STAGES OF BRECCIATION

FIGURE 1.—Base of slide, Schultz area. Contact of slide and 19 bed marked by dashed line. Upper part of 19 bed broken but still in normal position.

FIGURE 2.—Glide plane of upper slide, Owl Creek basin.

FIGURE 3.—Flowage within upper slide mass, Owl Creek basin.

FIGURE 4.—Reversal of dip and beginning of brecciation in upper slide, Owl Creek basin.

FIGURE 5.—Intense brecciation within upper slide, Owl Creek basin.

FIGURE 6.—Predominant basinward dip of breccia blocks, upper slide, Owl Creek basin.

FIGURE 3.—SLIDE ZONES ALONG RIMS ON LOCAL SEDIMENTARY BASINS

from the top of the 19 unit to the base of the 5 unit (Fig. 1). A slide rests on a calcarenite base. This may be the top of the 19 bed or, if other calcarenites are present above the 19, a higher calcarenite may serve as the base. Slides in the upper part of the 7 zone are restricted to areas in which buried igneous knobs or organic structures maintained the depositional environment for a considerable thickness of sediment. In the southwestern part of the Owl Creek basin and in parts of the Schultz basin such conditions continued throughout the first half of Bonneterre time, and great thicknesses of breccia were formed. In the number 10 basin the breccia is restricted almost entirely to the 12 bed. This area did not receive basin-type sediment after early Bonneterre time, and the environment was not suitable for breccia formation during the deposition of the higher beds.

TABLE 1.—SIZE OF MAJOR SLIDE AREAS

Basin	Length (in feet)	Width (in feet)	Thickness (in feet)	Tons of Rock
	Avg.	Avg.	Avg.	
#10	10,500	2000	40	76,000,000
Owl Creek	16,000	800	70	81,000,000
Leadwood	14,000	1200	40	62,000,000
Schultz	14,000	1500	50	95,000,000

Facies relationships.—The slides occur in the zone of facies change from bank-type calcarenite and reef to fine-grained basin sediments. On the south side of the Owl Creek basin, breccia is exposed in mine workings for over 6000 feet (Fig. 5). Differences in lithology are mapped on opposite walls of the stope. Outside the mined area several hundred diamond-drill holes reveal a pattern of increasing amounts of calcarenite to the south and basin sediments to the north. In the stratigraphic zone in which the breccia occurs complete sections of calcarenite and basin sediment are present within a few hundred feet of each other (Fig. 6).

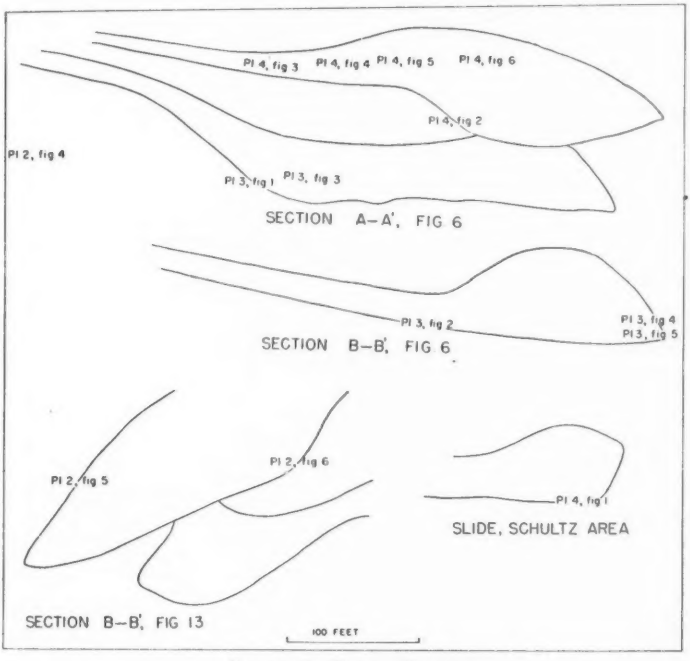

FIGURE 4.—KEY TO PLATES

At the eastern end of the area shown in Figure 5 the breccia lies stratigraphically above that part of the section for which the bank and basin facies boundaries are shown (Fig. 6). In this area lithologies of beds enclosing the higher

FIGURE 3.—SLIDE ZONES ALONG RIMS ON LOCAL SEDIMENTARY BASINS

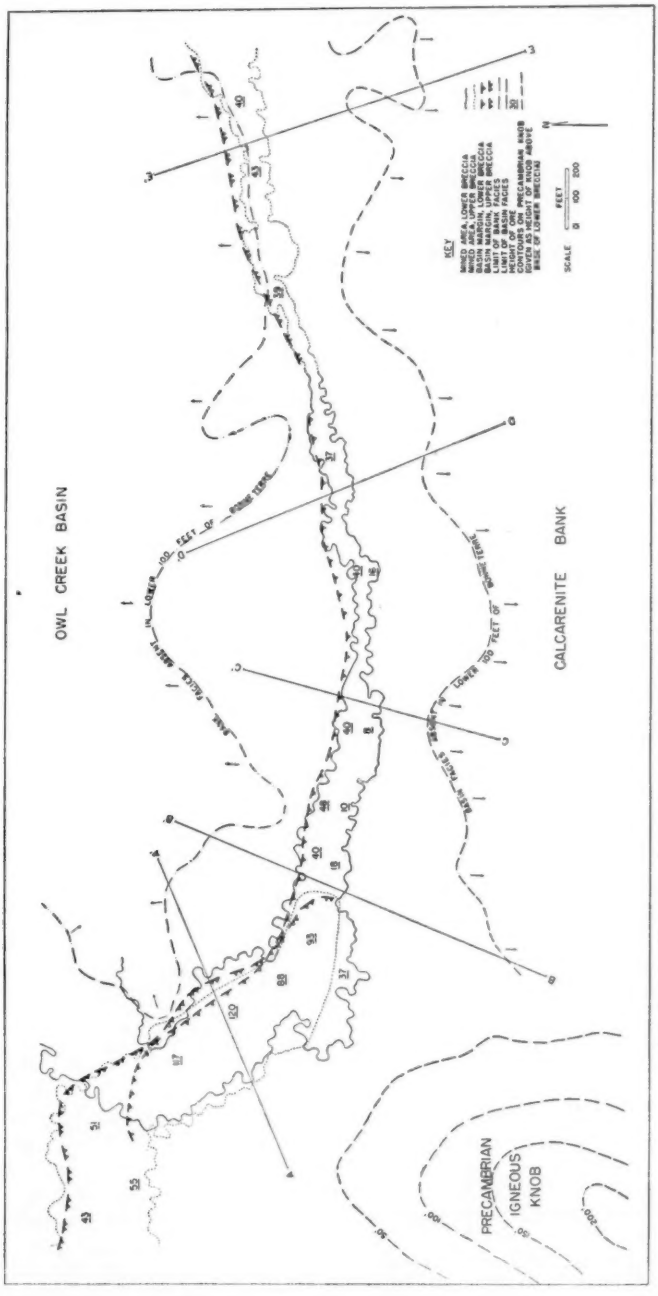

FIGURE 5.—BRECCIA ORE BODY AND FACIES RELATIONSHIPS, SOUTH RIM, OWL CREEK BASIN

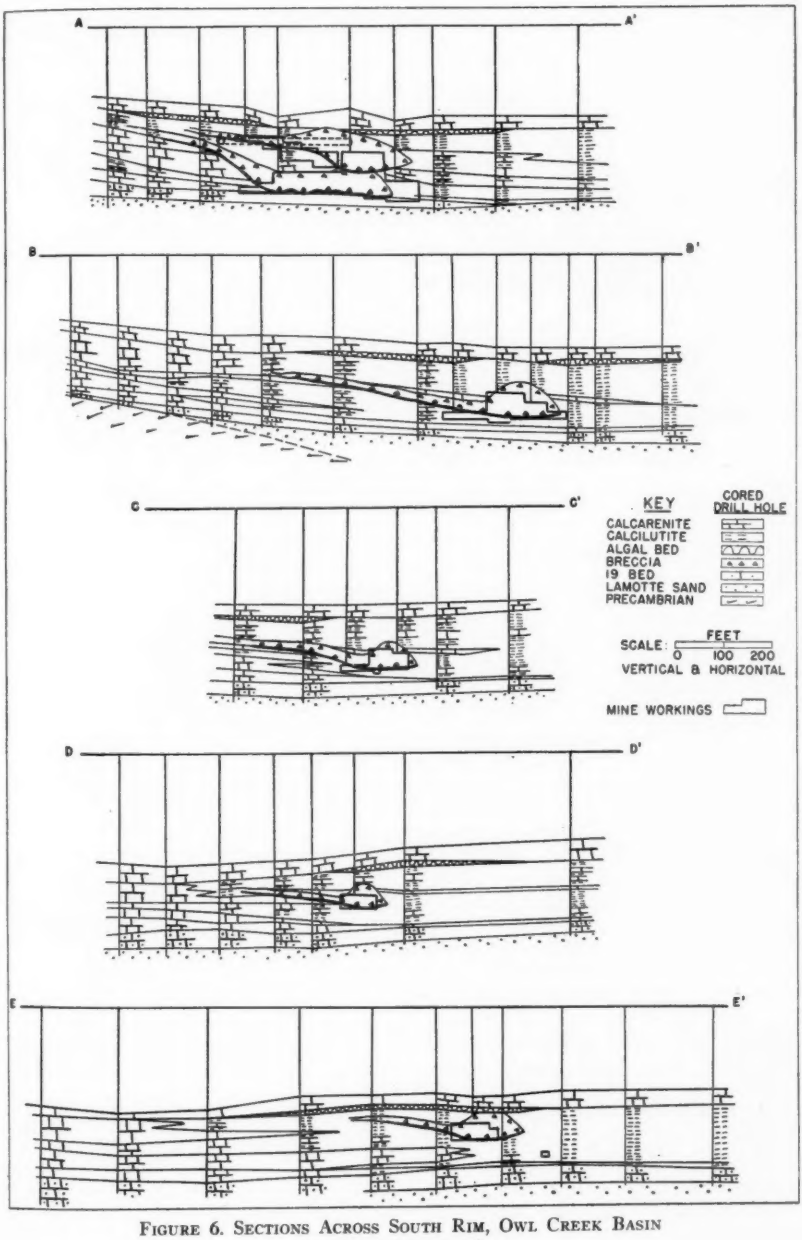

FIGURE 6. SECTIONS ACROSS SOUTH RIM, OWL CREEK BASIN
See Figure 5 for location

breccia indicate the bank facies has transgressed basinward.

Mine workings in the Schultz area are near Schultz basin, delineated by approximately 2000 diamond-drill holes, is in the zone of merging facies between platform calcarenites to

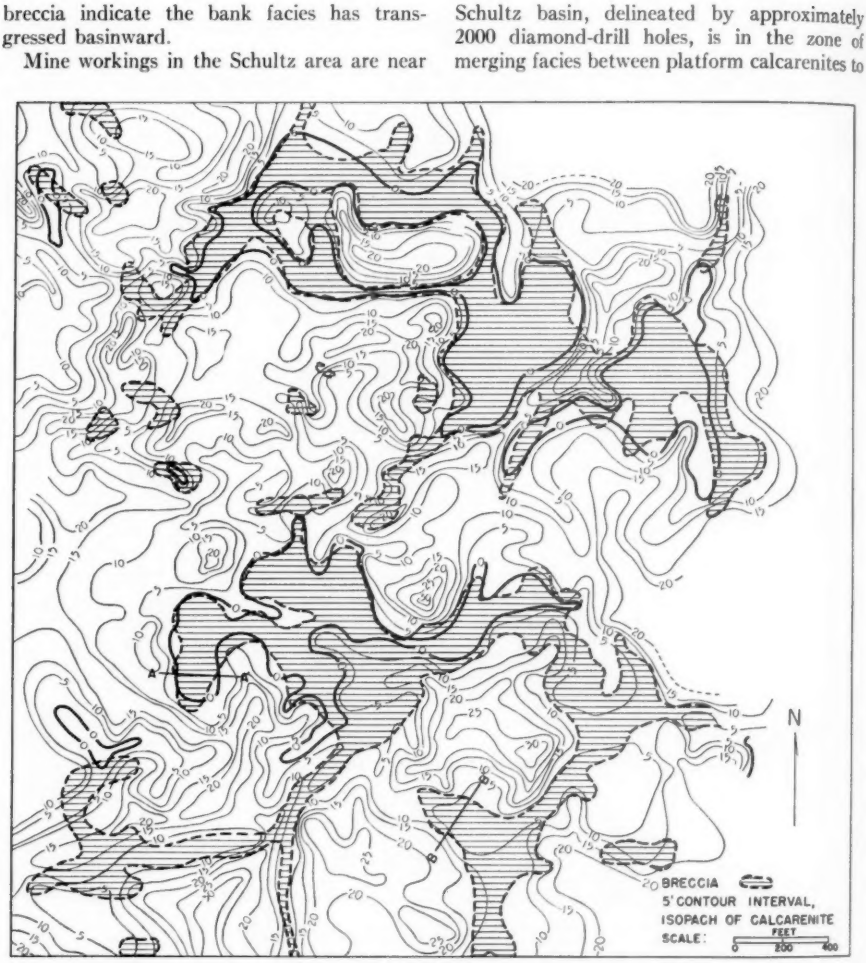

FIGURE 7.—LOWER BRECCIA ZONE AND ISOPACH OF CALCARENITE, SCHULTZ AREA
Areas shown in Figures 7, 8, and 11 are identical. (*See* Figure 3.)

the southwest end of a breccia zone over 3 miles long. The depositional pattern in this area was influenced by an igneous knob on the pre-Lamotte erosion surface. Bars of Lamotte sand and later of 19 and 15 bed calcarenites were formed adjacent to the knob. In the deeper waters between the bars and in the basin north and west of the knob, calcilutites, calcisiltites, and shales were deposited. The extremes of local depositional environments are shown on an isopach map of 15 bed calcarenite of the area (Fig. 7).

The great slide belt on the edge of the the southeast and basin sediments to the northwest. Over much of this belt slides are restricted to the lower part of the 12 bed, with the breccia base resting on 15 or 19 bed calcarenite. Near and over the buried igneous knob the fluctuating bank and basin environment was maintained into 7 bed time, and zones of breccia occur up to 120 feet above the basal calcarenites (Fig. 8).

Contact features.—The contact of the base of a slide with underlying beds may take several different forms varying with position on the structure. The contact may represent a "simulated unconformity" as described by Newell

et al. (1953, p. 86) where the mass cuts across underlying beds (Pl. 3, fig. 1; Pl. 4, fig. 2); a concordant relationship between slide and underlying units where movement has been the breccia may buckle and fold. Such drag folds are common where a thick bed of incompetent carbonate is near the base of the disturbed zone.

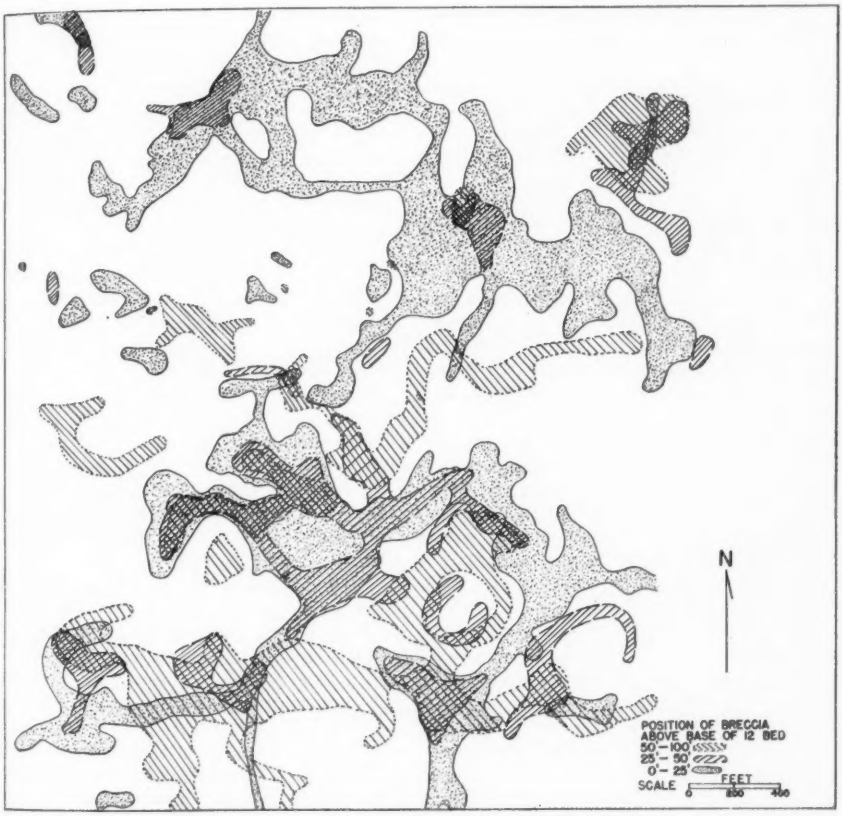

FIGURE 8.—SUCCESSIVE BRECCIA BODIES, SCHULTZ AREA
Areas shown in Figures 7, 8, and 11 are identical.

along a bedding plane (Pl. 4, fig. 1); or a gradational zone near the toe of the slide in which the base of the disturbed beds is marked by crumpling and drag folding (Pl. 3, figs. 4, 5).

The base of the slide, hereafter referred to as the glide plane, represents a concave, arcuate surface conforming to the basin outline. The glide plane may dip up to 60° on the flank of the calcarenite ridge. When viewed in limited exposure the structure may resemble a thrust fault. Basinward dip of the glide plane flattens gradually until it passes onto a bedding plane. Near the basinward margin of the breccia mass (the toe of the slide), beds in the basal part of

Slickensides are common on the base of the breccia mass where the glide plane is at a low angle or parallel to bedding (Plate 3, fig. 2). They have not been seen where breccia cuts across bedding at a steep angle. Slickensides are consistent in direction of strike over considerable areas and are in agreement with inferred direction of movement of the slide.

Wherever observed, the beds beneath the breccia body show little disturbance except for two types of features. Along the narrow zone marking the change of dip as the glide plane transgressed from a crosscutting to a bedding-plane relationship, beds below the breccia may

show considerable crumpling and slight brecciation. This condition varies with lithology of the underlying beds. The upper surface of the beds below a glide plane in places shows angular pits or depressions that appear to be the sites from which blocks were plucked or gouged during movement of the overlying sediment. A thick, competent bed below the glide plane may be fragmented, and tabular slabs may have moved short distances (Pl. 4, fig. 1). Plucked and transported blocks have been noted only at the top of the 19 bed at the base of thick breccia zones.

The upper surface of the slide dips away from the calcarenite bank at a slightly flatter angle than the base, resulting in a gradual thickening of the mass toward the basin. When the glide plane becomes concordant with bedding, dip of the upper contact is sharply reversed; the mass is abruptly thickened; and deformation becomes more intense (Pl. 4, figs. 3–6). Brecciation involving breaking of beds and dislocation of fragments succeeds intrastratal gliding. This greatly expanded and disturbed zone passes into undeformed beds on the basinward side by underthrusting and drag folding (Pl. 3, figs. 4–5). The reversal of dip of the upper surface of the slide forms a trough which is filled with undisturbed sediment. The axis of this trough can be traced for long distances in mine workings.

Within the trough and also overlapping the upper surface of the breccia on the basinward margin is a thin bed of shale and shaly carbonate which shows no disturbance. This bed, recognizable in mine workings and in drill core, is believed to represent resumption of normal deposition after the period of brecciation. Overlying the shale unit with conformable relationships are undisturbed beds of calcarenite and shaly calcilutite.

Internal structures.—Degree of brecciation varies with position in the mass, with slope relationships, and with lithology. The most intensely broken zone is at the toe, where underthrusting against basin sediments has fractured and displaced competent beds of calcilutite and caused incompetent beds to flow freely around the stronger fragments. Small thrust faults, thoroughly healed by flow of incompetent carbonate, are marked by alignment of calcilutite blocks. Within and adjacent to thrust zones brecciation is so intense that breccia blocks cannot be traced to individual unbroken beds. The zones of thrust faulting pass into drag folds. The folding involves movement along bedding planes accompanied by development of slickensides on bedding plane surfaces. Rock within the folded bed may be completely shattered.

Both types of internal structures—thrust faults and drag folds—parallel the basinward margin of the slide. Along the basinward side of the brecciated zone, folds are asymmetrical and overturned; axial planes dip 20°–50° toward the basin. Fractured and folded units can be seen in plan view on the roof of mine workings. Individual structures may extend for hundreds of feet of strike length. Within the greatly thickened toe of the slide but away from the basinward margin folds are more nearly symmetrical and axial planes are nearly vertical.

Successive exposures across the upper part of the toe of a slide are shown in Plate 4. Beds on the footwall side (Pl. 4, fig. 3) show a gentle basinward dip and slight disturbance. Little brecciation is apparent, but discontinuous units, pinching and swelling of beds, and minor internal thrust planes are common. Reversal of dip, at first gentle (Pl. 4, fig. 4), is followed by abrupt steepening and strong brecciation (Pl. 4, fig. 5). Within the breccia mass, blocks are oriented in random positions. Near the basinward margin of the slide, brecciation is intense, but over-all dip of blocks is toward the basin (Pl. 4, fig. 6).

Multiple breccia bodies.—Each major slide zone is composed of many individual slides. All slides within a zone can be referred to a few stratigraphic horizons. A breccia mass at a given stratigraphic position may be extensive or may consist of several distinct bodies separated by undisturbed beds. Breccia bodies superimposed one above the other may intersect or may be separated by unbroken beds. Position and trend of each breccia mass is controlled by bank and basin facies relationships at the stratigraphic level of the slide, consequently, successive breccia bodies may vary considerably in lateral position and trend.

Intersections result from a stratigraphically higher slide truncating the greatly thickened toe of a lower breccia zone (Fig. 6, section *AA'*). Great thicknesses of continuous breccia occur only at such intersections. In such cases, the upper broken zone may differ in strike from the lower breccia mass. Each level of breccia represents a distinct unit. If the breccia is mineralized, each zone mines as a separate mining level for most of its length; at the intersection mineralization may extend throughout the breccia mass, giving ore bodies of great thickness (Fig. 5).

At intersections, the glide plane of the upper

slide is difficult to recognize because of its similarity to internal structures. The outlines of individual bodies can be defined by lateral position at successive mine levels of such structural features as the steeply inclined glide plane, trough, and reversal of dip.

Mode of Formation of the Breccia

General factors.—Three general factors contributing to formation of the slides are the character of the depositional surface, differential compaction, and the effect of water in the sediment.

SLOPE OF DEPOSITIONAL SURFACE: The southwesterly regional dip within the main productive area (a reversal of original dip) was superimposed on the district by faulting. Attempts at restoration of areas of several square miles extent to their original positions give the best results with a regional dip of about a half degree to the northeast.

As described earlier, regional dip of the depositional surface in early Bonneterre time was complicated by the development of sand bars and banks. These bars and banks, locally called ridges, provided restricted zones along their flanks in which sediments were deposited on surfaces appreciably steeper than regional dip of the shelf area. Within the mine workings dips as steep as 25° are exposed (Pl. 2, fig 3). On the flanks of the larger depositional ridges dips of 8°–10° may be exposed for several hundred feet downslope and for thousands of feet along the strike. Although the observed dips may be due in part to differential compaction of sediments, the existence of an original dip of several degrees on the flanks of individual carbonate sand ridges is well established.

ROLE OF DIFFERENTIAL COMPACTION: The profound facies differences seen in the lower Bonneterre undoubtedly were reflected in differences in behavior of sediments in early diagenetic history, particularly in degree of compaction. The dominant features influencing deposition in early Bonneterre time was the carbonate sand banks and the extensive algal reefs growing over them. Depositional basins between sand banks received large amounts of lime mud, lime silt, and argillaceous mud. Terzaghi (1940, p. 88) showed that calcareous muds may undergo almost as much compaction as argillaceous muds. Carbonate sands of the banks and the rigid reef framework were probably only slightly compacted; the fine-grained basin sediments probably were strongly compacted. Compaction, beginning in each layer of mud as soon as it was buried by the next succeeding layer, would decrease gradually as basin sediments passed by intertonguing and intermixing into the bank facies. The gradation in degree of compaction would most strongly influence attitude of beds on the flanks of the depositional highs, and at these positions original dips would be greatly steepened. The steepening of dips in the zone of facies change would continue until compaction was essentially completed, although most of the steepening would have taken place under a relatively thin load of sediment.

FLUID MIGRATION: Compaction of sediments for the most part results from squeezing out of connate water. Part of the water may have escaped directly upward into the basins; impermeable layers of calcilutite or shale may have forced some water to migrate laterally.

The fluids that were forced to move along bedding planes and laterally through permeable beds migrated up dip on the flanks on the depositional ridges. As a result of this migration a great volume of water may have passed through restricted zones along the edges of large basins.

Origin of the breccia.—From the form and structure of the breccia masses and their relationship to the calcarenite bank sediments on the one side and basin sediments on the other, it is apparent that they formed by slump or gliding of unconsolidated sediments in the zone of transition from bank to basin facies. No other hypothesis satisfies the observed form and structure, the repeated occurrence at successively higher stratigraphic positions, the close relationship to particular sedimentary environments, and the lack of any relationship to known fault patterns of the district. As pointed out earlier, glide surfaces viewed in limited exposure may have the appearance of thrust faults. When considered in relation to the other elements of the breccia masses as exposed in mine workings, it is evident that they cannot be tectonically induced thrust faults.

The sequence of events giving rise to the breccia zones may be summarized as follows. Normal deposition of the lower part of the Bonneterre Formation formed an irregular pattern of calcarenite banks and local basins. Sediments over the banks were within the zone of water agitation (the *undaform* as described by Rich, 1951, p. 1–20), and silts and muds were winnowed out and swept into nearby basins. Minor changes in water level or depth of wave base resulted in interfingering of bank and basin sediments as one or the other achieved tem-

porary dominance. Although the surface of deposition lay above wave base over the banks and below wave base in the basins, only a relatively small difference in depth of water existed between the two environments. The sediment-water interface was essentially flat over the banks and nearly flat in the basins but was inclined basinward at a gentle slope along the flanks of the calcarenite ridges. With continued deposition the weight of overlying sediments forced the connate water of the basin sediments to migrate, resulting in compaction and thinning of the fine-grained units. Within the coarse-grained calcarenites large volumes of water could be retained in the interstices between closely packed grains; weight of overlying sediments would result in tighter packing of grains but relatively little decrease in volume. As the process continued, dip of beds on the flanks of the calcarenite ridges became so steep that the unconsolidated sediments slid downslope under the impetus of gravity (Fig. 9).

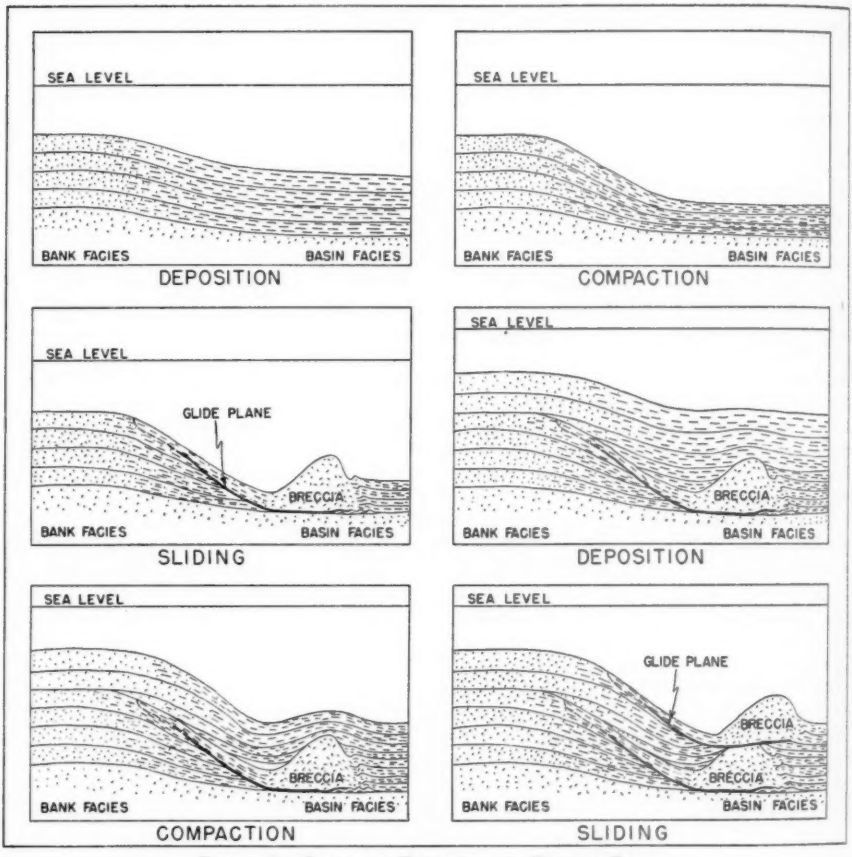

FIGURE 9.—STAGES IN FORMATION OF BRECCIA BODIES

The initial movement, involving only a thin veneer of unconsolidated and incompetent sediment, took place by sliding and flowage. Movement may have been aided by water migrating up the flanks of ridges. Frictional drag on underlying beds as the mass moved downslope set successively older beds in motion, increasing the thickness of the slide and the volume of moving material. The more deeply buried beds affected by the slide were lithified enough to exert a gouging action on underlying material, enabling the mass to cut downward.

The downward cutting was localized by a zone of fracturing that formed during compaction. While the calcilutite and shale beds were decreasing in thickness, the less compactible units were being lowered at differential rates, depending on proportions of the two types of material. Fracturing occurred in those units which could not adjust easily to differential settling. The base of the sliding mass in higher sediments cut down rapidly through the fractured zone to an underlying unbroken calcarenite bed. These processes, believed important in localizing the glide plane, are shown diagrammatically (Fig. 10).

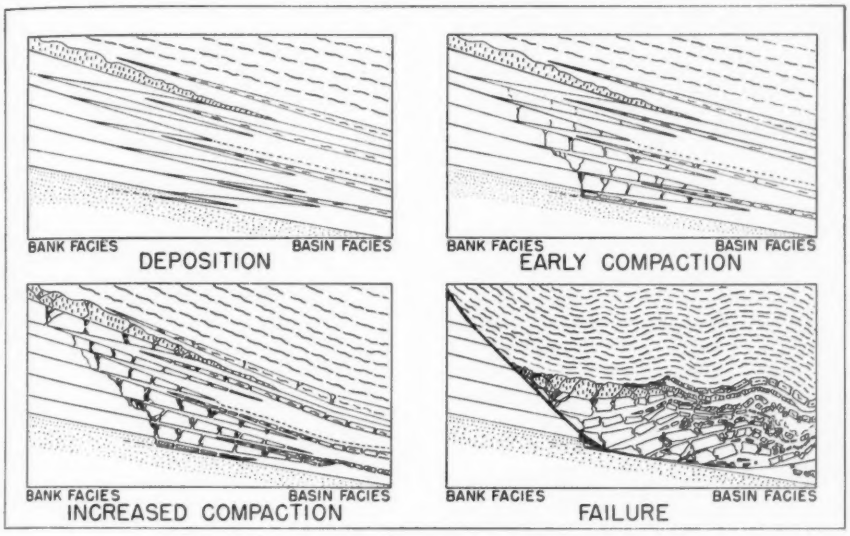

FIGURE 10.—STAGES IN LOCALIZATION OF GLIDE PLANE

The sliding mass, after reaching the foot of the slope, moved a short distance into the basin where its momentum was quickly checked by the buttress of basin sediments. The mass of material, forced to stop, buckled and arched to form a ridge of breccia at the base of the slope.

The point of origin of the slide is open to question. The movement of material downslope and withdrawal of support for beds farther upslope may have set sediments in motion higher on the ridge. The effects of the slide would advance headward up dip, although the slide itself was moving downslope.

Following the initial slide the breccia ridge was buried as normal deposition continued. If the bank and basin facies relationships remained unchanged with continued deposition, the process would be repeated when a sufficient thickness of sediment was deposited and compaction again resulted in oversteepening of dip (Fig. 9). Two slump episodes along a basin margin are common; three are not unusual; and on the southwest margin of the Owl Creek Basin at least five distinct periods of slump have been recognized.

Individual features within slides vary with local relationships of slope and lithology. Glide planes differ considerably in detail. The shape of the trace of the glide plane when viewed in section can be seen to have been influenced directly by facies relationships. In section *AA'* of Figure 6 the lower breccia mass cut downward until it reached the coherent surface of the 19 bed. The broad flat breccia mass with a small pile-up at the outer margin is a result of a steep slope and a thin zone of basin sediment. The upper breccia mass moved on a gentle slope over calcarenite beds. When the mass reached a zone of predominantly basin-type sediment, it gouged downward abruptly (measured angle of 48°) and flattened out only when it cut into the underlying mass of breccia. The heterogeneity of the breccia and the absence of bedding planes to facilitate sliding undoubtedly helped check the downward movement. In section *BB'* of Figure 6 the slide occurred on a long gentle slope, and momentum was halted at the foot of the slope.

Slickensides and plucking of blocks from underlying beds indicate that the slides cut downward into sediments of some degree of coherence. Slickensides would not form by flow of incoherent material alone. Their common occurrence indicates that beds at the base of the slide were at least partially lithified. Plucking of blocks from underlying beds could be done only when sediments at the base of the slide were coherent enough to gouge into the underlying unit and enclose the disrupted block. The coarse fragmentation common in the lower parts of breccia zones also indicates disruption of coherent material as contrasted with easy movement along bedding planes in incoherent sediment.

The zone of slight brecciation below the glide plane along the line at which the glide plane passes onto bedding attests to the momentum of the mass. In some places where the glide plane is not pronounced, it is difficult to define the base of the slide and distinguish between breccia in the slide and breccia below. This restricted zone of breccia below the slide suggests that the moving sediments may have reached the foot of the slope with considerable momentum and that part of the force of the slide was directed downward before the mass flattened out along bedding.

The disturbance created by the slide and the impact of its plowing into unconsolidated basin sediments put in suspension large amounts of fine sediment. The finely laminated shale and shaly calcilutite overlying the breccia and adjacent undisturbed beds are believed to be due to redeposition of this material.

Factors of Breccia Thickness and Compaction

Calculated data.—Extensive underground exposures, detailed three-dimensional mapping, and diamond drilling on approximately 100-foot centers permit measurement of components of slides on the south rim of the Owl Creek basin. Statistics, measured along sections shown in Figure 6, are listed in Table 2.

The slides at the position of section *AA'*, Figure 6, are somewhat anomalous due to the influence of the towering Precambrian knob (Fig. 5). To the east, as shown on sections *BB'* to *EE'*, inclusive, the slides are directly related to a calcarenite ridge built outward from the knob rather than to the knob itself. Sections *BB'*, *CC'*, and *DD'*, for which the glide plane can be traced continuously in mine workings, show mutually similar relationships of angle of slope and thickness of breccia.

In all sections except the lower slide of section *AA'* the height of the calcarenite ridge above the basin ranges from 30 to 40 feet. Present slope of the plane, believed to represent the surface of the ridge between the point of initiation of the slide and the correlative basin surface, is approximately 8°. As noted in Table 2, the top of the 19 bed is used as a reference plane. Irregularities on the 19 surface might modify the measured distances, but the resulting error is small and would not affect results materially. The measured slope would include

TABLE 2.—CALCULATED FACTORS OF SLIDE ZONE

Cross-section and breccia zone	Thickness of bank sediments, from top of bed 19 to sediment-water interface (in feet)	Thickness of basin sediments, from top of bed 19 to sediment-water interface* (in feet)	Vertical difference, bank to basin (in feet)	Calculated slope (in per cent)	Depth of gouging (in feet)	Thickness of breccia pile-up (in feet)
AA' (Lower slide)	118	38	80	16.0	38	72
AA' (Upper slide)	136	103	33	7.5	44	86
BB' (Lower slide)	92	58	34	6.8	32	68
CC' (Lower slide)	94	58	36	7.1	32	56
DD' (Lower slide)	88	64	24	7.5	30	52
EE' (Upper slide)	122	92	30	8.8	34	58
Average, All	108	69	39	8.9	35	65
Average, Lower slide	98	54	43	9.3	33	62
Average, Upper slides	129	97	31	8.1	39	72

* Thickness of basin sediments from top of 19 to sediment-water interface is taken as thickness of sediment from top of 19 to a persistent undeformed calcarenite bed overlying calcilutites and shales and abutting against breccia pile-up. This bed was not involved in slumping and would give the maximum thickness of basin sediments. Possibly the section as given could be a few feet less; it could not be greater.

effects of postbrecciation compaction and therefore would represent a maximum angle.

The depth to which the slide gouged into basin sediments is fairly uniform on all sections and is roughly equal to the height of the ridge above the basin. Thickness of the breccia pile-up at the foot of the slope is approximately twice the depth to which the slide gouged into basin sediments. These relationships appear to be fairly constant in the slides on the south side of the Owl Creek basin.

The lower slide shown in section *AA'* is anomalous in thickness of calcarenite, in thinness of basin sediment, and in angle of slope. The thick deposit of calcarenite is typical of the knob environment. The thin section of basin sediment in this area is confirmed by isopach maps of stratigraphic units. This part of the basin received no 19 zone sediments; the 12 bed rests directly on the Lamotte Sandstone. That part of the 12 zone involved in the lower slide is thinner than normal. The thin section of basin sediment disturbed by the slide, as well as the steep slope, permitted the slide to move well out into the basin before its momentum was checked.

The calculations on height and depth (Table 2) are supported by observed lithologic relationships. The top of the breccia ridge stood at about the same height as the calcarenite ridge; a lagoonal area separated the two ridges. In section *EE'* of Figure 6, drilling indicates a flat-topped breccia mass, suggesting that the top of the ridge may have been eroded. Algal beds growing over the upper breccia (sections *AA'*, *EE'*) indicate the change from basin to shallow-water environment over the arched-up mass.

Effect of character of sediment.—Within a breccia zone, relative competency of the material varies considerably due to different types of sediment involved and different degrees of compaction. At any given stratigraphic position (depth of burial) the fine-grained, homogeneous calcilutite appears to be the most competent material present during early diagenesis. Invariably, it shows the least inclination to flow and the strongest tendency to maintain its identity by breaking into coherent fragments. In contact with mudstones, the calcilutite may be unbroken and undeformed whereas the argillaceous material may be crumpled and broken from intrastratal movement. All coarser-grained clastic carbonates are less competent than the calcilutite; within the breccia bodies the calcilutite has broken into angular fragments, around which the calcarenite has adjusted by flow.

Newell *et al.* (1953, p. 159), suggest that compaction of basin sediments of the Capitan Reef area was essentially completed within 100 feet of the sea floor. The Lead Belt breccias indicate that for relatively pure carbonate muds, most of the compaction takes place with less than half this thickness of sediment. The calcilutite blocks in the lower parts of slides represent material which was buried by 30–40 feet of sediment at the time of brecciation. The blocks form the most striking examples of brecciation and lie at all conceivable attitudes. Thinning due to compaction would take place in the vertical plane. Consequently, blocks from a given bed thrust into random positions should show differences in thickness according to attitude of the block if appreciable compaction occurred after brecciation. This is not the case; fragments of a given bed, regardless of attitude, have essentially the same thickness.

The coherence attained by the fine-grained carbonate at shallow depths may be due to rapid crystallization as well as to simple compaction. Fineness of grain and homogeneity of material would lead to early lithification while coarser-grained sediments were still unconsolidated.

The figures (Table 2) for the Owl Creek basin slides reflect certain conditions of slope and degree of compaction. Under different environmental conditions movement of sediments could be initiated which would affect only slightly compacted, nearly fluid materials. Mass movements of sediments on the ocean floor may vary from downward-cutting slides of the type described to submarine mudflows. The more fluid materials would move on very low-angle slopes, do little downward cutting, and spread widely with little pile-up at the foot of the slope. Individual exposures in some mined areas are suggestive of mud flows, but because of the difficulty of recognizing such material in drill core, no such areas have been delineated.

A type of movement which appears to mark a stage intermediate between the downward-cutting slides and mud flows is represented by the fourth period of slump in the five slump episodes of the Owl Creek basin. The disturbed zone lies southwest of the area shown in Figure 5 and over the buried igneous knob. Limited exposures in mine workings show parallel folds of beds of calcilutite and shaly calcarenite. The folds have a wave length of approximately 30 feet and an amplitude of 10 feet. The base of the folds is marked by an indistinct glide plane. The disturbed zone dies out gradually. Similar disturbed zones in which the undulating beds extend laterally for several hundred feet are

common in the lower slide in parts of the Schultz area.

Inferred Character of Submarine Slides

Submarine slides of the Bonneterre Formation, as determined from adequate three-dimensional data, represent intermittent mass movements of individual blocks of unconsolidated sediment. The slides occurred repeatedly throughout a considerable stratigraphic thickness in restricted environmental situations. Within the area shown in Figure 5 the breccia long was regarded as a single slide. Recognition of component parts of slides and tracing of glide planes within mine workings revealed intersecting slides of different ages.

If in the southeast Missouri district the present erosion surface were a few hundred feet lower, one might find large areas of breccia exposed over a linear distance of approximately 10 miles. All would be encompassed in a stratigraphic interval of 200 feet. With incomplete exposure all might be attributed to a single period of slump. The repeated occurrence of slides in individual isolated areas as defined in the Lead District suggests that estimation of volumes of breccia made on the basis of surface exposure should be done with caution.

Mineralized Breccia Structures

Character of Breccia Mineralization

General features.—The dolomite breccia is an excellent host rock for ore; it has provided many small- to medium-sized and several large ore bodies of better than average grade. Mineralization of the shale breccia is rare; where it occurs it is not of ore grade. Descriptions of breccia ore which follow apply only to the dolomite type.

The dolomite breccia in places carries lead, zinc, and iron sulfides. Although throughout the district lead is far more abundant than zinc, in the basin type, lithology of the breccia and associated beds zinc may be more abundant locally than lead. Zinc is an important constituent of ores in parts of the Owl Creek and Leadwood breccia areas, but it is absent in the Schultz breccia. On the south side of the Owl Creek basin zinc mineralization extends hundreds of feet beyond lead mineralization. Distribution of the zinc is one of the still undeciphered problems of mineral zoning.

Galena in the breccia ore occurs as irregular veinlets around breccia blocks, as numerous scattered blebs and small crystals in the breccia matrix, and as irregular massive replacements of the matrix. Distribution of the galena is erratic, although the better grade is concentrated in the zones of chaotic brecciation.

Ore in breccia is always associated with underlying bedded ore at one of the favorable lithologic contacts. The breccia zone above a mineralized contact gives great thickness to an ore body which otherwise might have no more than minimum mining height. Fragmentation of impermeable beds during brecciation as well as secondary fracturing appears to have permitted easy movement of mineralizing solutions above the bedded ore zone. Bedded ore may extend laterally far beyond the limits of the breccia body, but breccia ore is not found beyond the limits of bedded ore.

Mineralization in the breccia is restricted to the piled-up mass at the basin margin. On the basinward side ore extends out to the limits of the disturbed area. The sloping up-dip part of the slide rarely is mineralized. The glide plane may be seen passing into the back of the stope, but mineralization continues below the glide plane as contact ore only.

Role of secondary fractures.—Postlithification fracturing associated with the breccia ore bodies consists of two apparently unrelated types. One type occurs as zones of northwest-trending faults that are part of the general fault pattern of the district. Except for one major fault which cuts across the eastern edge of the Owl Creek basin, most of the faults show displacement of less than 10 feet. The faults show no apparent relationship to the breccia bodies and cut through breccia and undisturbed beds alike.

The other type consists of tight knife-edge fractures that occur in swarms and are found only in the breccia. The fractures are of short extent both vertically and horizontally; they terminate downward near the base of the slide on shaly bedding planes; they do not penetrate the underlying calcarenites. They show many diverse trends, and many roughly parallel the outer edge of the breccia mass. The fractures pass through breccia fragments and matrix indiscriminately; they may cross healed thrust zones within the breccia without deflection.

The latter type of fractures shows no relationship to fault patterns of the district. The diversity of strike, the discontinuity, the confinement to breccia zones, and the attitude of the fractures relative to adjacent undisturbed beds suggest that they are purely local phenomena. They are attributed to interruption of

tectonically induced bedding-plane movement by the massive structureless body of breccia. Many of the irregular fractures bear veinlets of galena and pyrite. They may have aided mineralization by providing channels and openings through which the ore solutions could move upward from underlying bedding planes.

Breccia Ore Bodies

Owl Creek basin.—The mineralized breccia bodies represent the slides which cut down to a bedding-plane contact that later was mineralized or which intersected other breccia bodies that cut a favorable contact. The mineralized portion of the Owl Creek breccia, part of which is shown in Figure 5, is a continuous ore zone over 6000 feet long that encompasses three distinct breccia bodies. The lower breccia body cut down to the 19 bed, and extensive mineralization occurred at the contact of these beds. Galena was disseminated in the upper part of the 19 bed. The entire piled-up mass of breccia is mineralized to the basinward margin of the breccia. A cross section of mine workings shows a typical breccia stope pattern. On the footwall side, below the glide plane, contact ore is mined in interbedded calcarenites and shaly calcilutites. The low back of this stope intersected the basinward-plunging glide plane. Where the glide plane flattened out along bedding and the upper surface of the slide turned sharply upward, height of mining increased; most of the breccia ridge is exposed. The mining floor was carried a few feet below the glide plane. The steep reverse-dip side of the top of the breccia is marked by a massive zone of lead and iron sulfide (Pl. 2, fig. 5), permitting easy recognition of the inclined breccia top in high mine workings.

The eastern edge of the lower Owl Creek slide approaches but does not actually appear to intersect a higher breccia zone (Fig. 5). If an intersection did exist, it has been destroyed in mining. Mineralization is continuous between the two breccia bodies, although the base of mineralization of the higher breccia is approximately 20 feet above base of the lower.

The western part of the lower slide is truncated by a higher breccia zone. In the area where the slides intersect, mineralization is continuous from the base of the lower breccia to the top of the upper, giving an ore zone up to 130 feet thick. At the area of intersection, the entire breccia zone is mined as a single ore body; away from the area of intersection each breccia mass constitutes a separate mine level. This composite ore body has produced several million tons of ore.

Stratigraphically higher breccia bodies southwest of the area (shown in Fig. 5) form separate ore bodies. The higher breccia zones do not intersect the lower zones, and no ore connection between the two groups is known.

The Schultz area.—Ore production from the Schultz breccia area has come from a large number of individual slides. The lower disturbed zone, resting on the top of the 19 bed, has produced approximately two-thirds of the breccia ore mined in this area. Distribution of the lower breccia which is related to minor calcarenite ridges is shown on Figure 11. The close coincidence of the breccia with areas of thin or no calcarenite and a similar close relationship of breccia to mined areas indicates the importance of small-scale facies changes in localizing ore bodies.

Successive breccia bodies are shown on Figure 8. For convenience these are grouped into a lower level which rests on the 19 bed, an intermediate level which corresponds to the middle of the 12 bed, and an upper level which embraces the upper part of the 12 bed and the lower part of the 7. The two upper breccia zones, roughly equal in areal extent, together have produced about one-third of the breccia ore mined in this area. Where the upper breccia bodies intersect lower breccia masses, mineralization is continuous to a height of 100 feet or more. The areas of high mining can be seen on Figure 8 as superimposed breccia bodies. Except where an upper breccia mass truncates a lower one, giving a very thick disturbed zone, each level of breccia ore is independent from other levels, and trend of stopes at different levels may vary considerably. The form of the piled-up breccia mass and the resulting outline of mine workings (Figs. 12 and 13) vary a great deal due to minor local facies differences influencing development of the slides.

Mineralization of all levels of breccia in the Schultz area is related to bedded ore from the contacts of 12 and 19 or 12 and 15 bed. The high levels of ore occur only where intersections of breccia provide access to the upper breccia zones. Although individual ore bodies may be small, combined production from breccia ore bodies of this area totals several million tons.

The Leadwood breccia.—Large tonnages of ore have been produced from the Leadwood breccia area, especially from the wide southern part of the zone (Fig. 3). The breccia zone is composed of many individual slides. Intersections of

slides are common, resulting in large and thick ore bodies, but many slides appear to be independent and form isolated ore bodies.

10 basin carry little or no ore, because the underlying beds do not carry ore.

The three western breccia areas contain

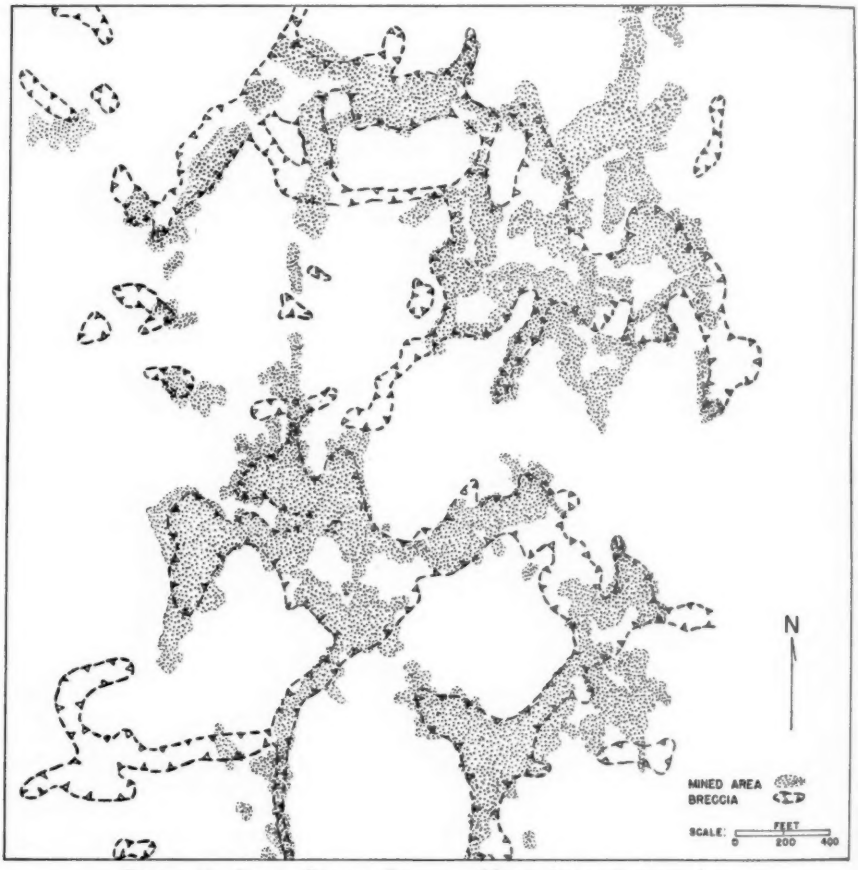

FIGURE 11.—LOWER BRECCIA ZONE AND MINED AREAS, SCHULTZ AREA
Areas shown in Figures 7, 8, and 11 are identical

Breccia mineralization is related to a wide-scale bedded-ore horizon. The contact ore is mined as large, low stopes, which increase to 60 feet or more in height in the breccia zones.

Importance of Breccia Zones as Ore Structures

Only a small part of the breccia in the district is ore-bearing, because its mineralization appears to depend on a direct connection with an underlying favorable ore horizon. Large areas of breccia such as that adjacent to the Number

numerous ore bodies from the top of the 19 up into the 7 zone and account for enormous tonnages of ore.

It is difficult to assess the importance of the breccia-type structure in terms of relative importance in production as compared with other sedimentary structures of the district. Many small individual ore bodies are of the breccia type. Many other small, mined-out ore bodies have a suitable environmental setting, but mining to the limits of the ore body has removed any breccia which may have been present.

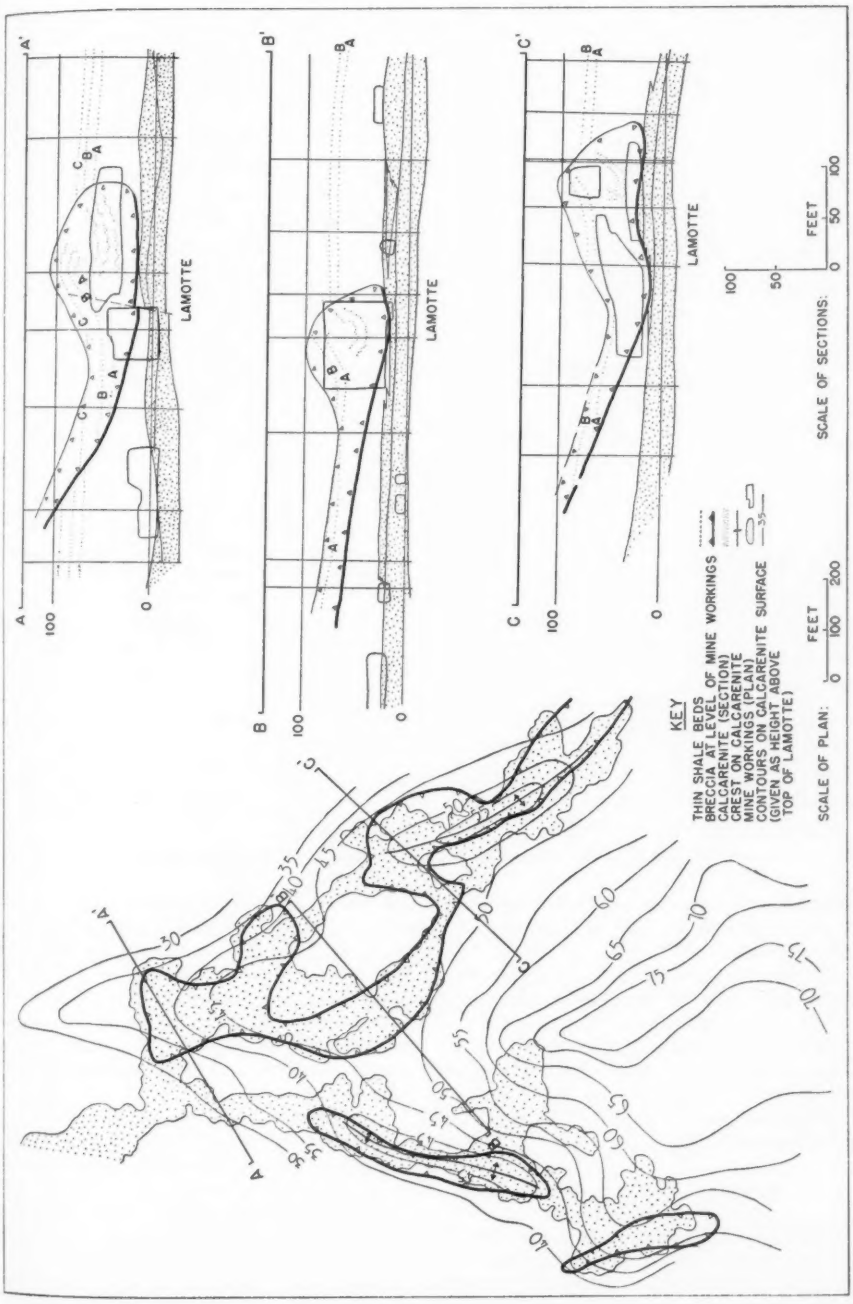

FIGURE 12.—BRECCIA ORE BODIES, SCHULTZ AREA
(*See* Figure 3.)

Breccia structures may have carried significant amounts of ore in stopes which were mined primarily on other types of structures. Throughout the western part of the district the presence of large breccia zones has resulted in increasing, in places severalfold, the vertical extent of mineralization.

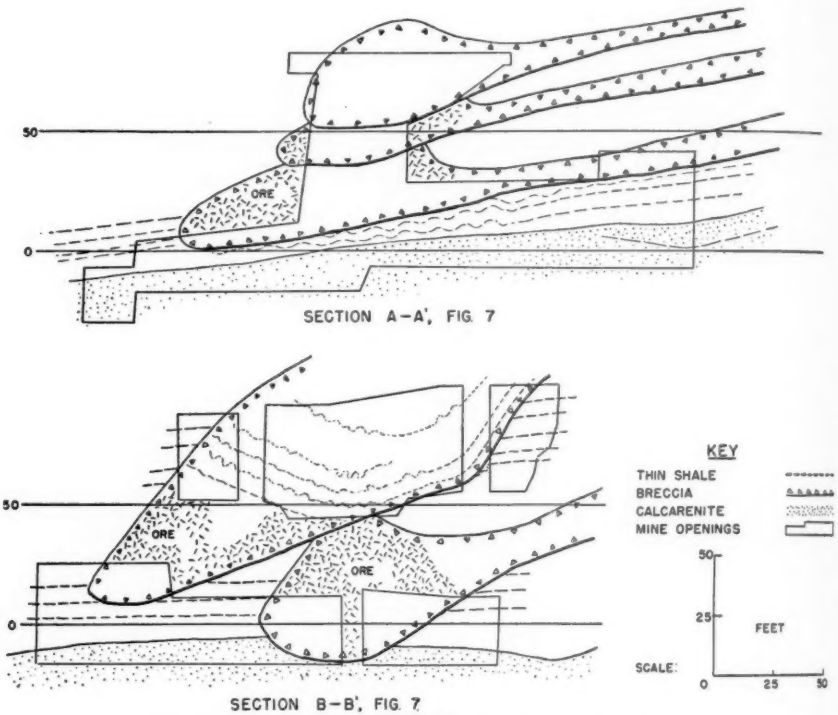

FIGURE 13.—SECTIONS ACROSS BRECCIA ORE BODIES, SCHULTZ AREA

At least 30 per cent of all production from the western part of the district came from breccia structures and related contact ore. A history of production of many millions of tons of ore from submarine slide zones in the southeast Missouri district requires that the slides be classed among the important ore structures of sedimentary rocks.

SUMMARY AND CONCLUSIONS

Lithology of the lower part of the Bonneterre Formation in the southeast Missouri lead district can be described in terms of four end-member types of sediment. These are argillaceous mud, sand-size sediment, recognizable algal deposits, and carbonate mud. Diagenetic changes have modified the original carbonate sediment, but the character of the original material can be determined by study of thin sections, hand specimens, gross sedimentary structures, and facies relationships. Sediments that were originally nearly uniform in type of material retain a considerable degree of uniformity after diagenesis; sediments composed of mixed sizes and types of sediment form rocks of varied texture because of the selective action of diagenetic processes.

Depositional highs composed mainly of calcarenite and *in situ* algal reefs formed prominent ridges, whereas intervening basins received deposits of argillaceous mud and carbonate silt and mud. This ridge and basin topography was the dominant control of deposition in early Bonneterre time.

Differential compaction of fine-grained basin deposits and relatively rigid calcarenites re-

sulted in oversteepening of dips in the gradational zone on the flanks of ridges and mass movement of unconsolidated sediments. The slides gouged downward into basin sediments. When momentum was checked by the buttress of basin sediments, the slide arched upward to form a breccia ridge at the foot of the slope. Calculations on slides indicate that a slide gouged into basin sediments to a depth approximately equal to the height of the ridge above the basin. Pile-up of breccia at the foot of the slope above the floor of the basin is nearly equal to the depth of gouging. Compaction of carbonate muds, probably aided by crystallization, was essentially completed under a load of 40 feet of sediment.

If the ridge and basin facies relationships remained unchanged, slides occurred repeatedly at frequent stratigraphic intervals. Slides of one age may truncate and overlap earlier slides, resulting in continuous zones of breccia up to 130 feet thick. The pattern of submarine slides in the Bonneterre Formation is one of relatively small individual movements of sediment that occur repeatedly throughout a considerable stratigraphic thickness in restricted sedimentary environments.

Where a breccia mass intersects an underlying contact favorable for mineralization of the bedded type, the entire breccia mass may be mineralized, giving ore bodies of considerable thickness. Intersecting breccia bodies may show continuous ore at the point of intersection, with mineralization extending throughout the nonoverlapping parts of each breccia mass. It is estimated that about 30 per cent of past production from the western part of the central productive area has come from breccia ore bodies. Production of many millions of tons of ore requires that submarine slide zones be recognized as an important type of ore structure in sedimentary rocks.

References Cited

Illing, Leslie V., 1954, Bahaman calcareous sands: Am. Assoc. Petroleum Geologists Bull., v. 38, p. 1–95

Newell, Norman D., Rigby, S. Keith, Fischer, Alfred G., Whiteman, A. J., Hickox, John E., and Bradley, John S., 1953, The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico: San Francisco, W. H. Freeman and Company, 236 p.

Ohle, E. L., and Brown, J. S., *Editors*, 1954, Geologic problems in the Southeast Missouri Lead District: Geol. Soc. America Bull., v. 65, p. 201–222, 935–936

Pettijohn, F. J., 1949, Sedimentary Rocks: New York, Harper and Brothers, 526 p.

Rich, John Lyon, 1951, Three critical environments of deposition, and criteria for recognition of rocks deposited in each of them: Geol. Soc. America Bull., v. 62, p. 1–20

Snyder, Frank G., and Emery, John A., 1956, Geology in development and mining, Southeast Missouri Lead Belt: Mining Engineering, v. 8, p. 1216–1224

Tarr, W. A., 1936, Origin of the Southeast Missouri Lead Deposits: Econ. Geology, v. 31, p. 712–754, 832–866

Terzaghi, Ruth D., 1940, Compaction of lime mud as a cause of secondary structure: Jour. Sed. Petrology, v. 10, p. 78–90

Wagner, R. E., 1947, "Lead Belt" geology: Mining and Metallurgy, v. 28, p. 366–368

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Short Notes

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WRENCH-FAULT TECTONICS: A DISCUSSION

BY JOHN C. MAXWELL AND DONALD U. WISE

Moody and Hill (1956) are to be commended for their analysis of strain patterns associated with wrench faults. Their paper has been the subject of much comment, particularly their theories relating wrench faults to a single meridional stress (Alberding, 1957) and the associated theory attributing many of the major tectonic features of the earth to a wrench fault origin. Although these theories are attractive simplifications, certain mechanical and geologic difficulties should be considered.

The theories propose that the earth's crust is broken into a number of blocks by great wrench faults resulting from north-south crustal compression. Eight directions of wrench faulting are postulated. These include two directions of first-order shearing arranged symmetrically about the north-south compression axis plus related second- and third-order shear directions (Fig. 19 of Moody and Hill's article). These eight shear directions are named after major fault trends, and parallel tectonic trends are suggested as examples of the same shear pattern in other parts of the world. Movements on the first-order shears cause local reorientation of the stresses. These new local stress environments produce second-order shears, and these in turn may reorient the stresses to cause third-order shears. Thus a corollary to the theory should be that second-order shears can result only from movement on first-order shears and that movement on third-order shears can result only from movement on both first- and second-order shears. In this manner a hierarchy of fault types is established with smaller and less important faults constituting the higher orders.

The general theory may be tested by application of this "hierarchy corollary" to specific cases. For example, one of the third-order directions is the New Zealand trend, with a lateral displacement which may reach 300 miles (Wellman, 1954); the other is the Texas lineament, interpreted by Moody and Hill as a large-scale left-lateral wrench fault. The Oca and Bartlett "second-order" lineaments are likewise commonly interpreted as wrench faults with lateral displacement measured in miles or perhaps tens of miles. If these named faults are truly second- and third-order shears, they must result *directly* from movement on related first- and second-order shears respectively. Furthermore, the movement on the first- and second-order shears should be essentially contemporaneous with, and exceed in magnitude, the movement on the resulting third-order shears. Thus the third-order New Zealand wrench fault should be the result of larger movement on an unrecognized second-order wrench fault which in turn was produced by even larger movement on a similarly unknown first-order wrench fault. A like relation to lower-order shears should be demonstrable for all shears of second and higher orders. For few if any cases can this be done, in spite of the fact that the lower-order shears should be larger and more prominent than the higher-order shears to which they give birth. The simple interpretation of a common origin for the earth's major wrench faults is therefore not acceptable unless a proper time and spatial relation can be demonstrated for a related group.

Alberding (1957) proposed such a system of related wrench faults for northern South America. He assigned first-order significance to the relatively minor Tigre, San Francisco, and Los Bajos faults, while the major displacements, of perhaps hundreds of kilometers, took place on "second-order" faults, El Pilar-Oca, and in an approximately *east-west* direction. If these faults can indeed be related in an order sequence, then the east-west system must be first order, the smaller faults related to them of higher orders, and the maximum compressive force must act more nearly east-west instead of in the postulated S. 15° E. direction.

Analysis of major wrench faulting might be approached in another way. Major shears in

the earth's crust may be generated either by a couple or by compression alone. Large shearing movements generated by a couple would in general be accommodated along a single wrench fault or a zone of parallel wrench faults. Shearing due to compression alone would be characterized by movement on a conjugate set of shears. The two types of shears have been designated respectively as simple shear and pure shear (De Sitter, 1956, p. 120). Moody and Hill treat only the conjugate fractures of pure shear. The alternative of simple shear by a couple also should be considered.

For differentiating between these two types of shear fractures the magnitude of the force normal to the shear plane is a helpful criterion. In simple shear the total force across the shear plane is an independent factor which can be highly variable. In pure shear, on the other hand, the force normal to the shear plane bears a direct relation to the sine of the angle of shear. For this angle to be equal to 30°, as postulated by the authors, the normal force across the shear would be one-half the primary compressive force. Thus, the mechanics of pure shear necessitate compression across the shear plane in *every* case, whereas the mechanics of simple shear have no such requirements.

The lack of brecciation on great thrust faults, generally regarded as examples of pure shear resulting from compression, is attributed to the strong normal forces which hold the walls together so tightly that the volume increase associated with brecciation cannot occur. If wrench faults are similarly the result of pure shear by great compression, then strong normal forces must exist across the plane in *every* case, and extensive brecciation along the fault plane would *not* be expected. The field evidence of wide zones of shattering and brecciation along most major wrench faults is a direct contradiction to these mechanical principles of pure shear. On the other hand, simple shear by a couple would predict variable normal forces, with resulting zones of widespread brecciation where such forces were small, and associated compressive features where the forces were large. The behavior of most wrench faults would seem to favor an origin by simple shear.

If major wrench faults are a result of simple shear rather than pure shear, the search for great compressive forces as the causal agent should give way to a search for mechanisms capable of shifting segments of the earth's crust by application of relatively local tangential forces. One such mechanism is subcrustal viscous flow, of which convection currents are a possible cause. Others might be envisioned, but at present the nature of the implied tangential forces is solely a matter of speculation.

Moody and Hill's theory of reorientation of stress along fault zones is certainly valid and will be most helpful in relating the various tectonic elements of a given region. However, at the present stage of knowledge, the extension of this theory into a simple unidirectional system of global wrench-fault tectonics seems unwarranted.

References Cited

Alberding, H., 1957, Application of principles of wrench-fault tectonics of Moody and Hill to Northern South America: Geol. Soc. America Bull., v. 68, p. 785–790

De Sitter, L. U., 1956, Structural geology: New York, McGraw-Hill Book Co., 552 p.

Moody, J. D., and Hill, M. J., 1956, Wrench-fault tectonics: Geol. Soc. America Bull., v. 67, p. 1207–1246

Wellman, H. W., 1954, Active transcurrent faulting in New Zealand (Abstract), Geol. Soc. America Bull., v. 65, p. 1322

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WRENCH-FAULT TECTONICS: A RESPONSE

BY J. D. MOODY AND M. J. HILL

We are very appreciative of the discussion by John C. Maxwell and Donald U. Wise of our paper entitled *Wrench-fault tectonics*. As is usual in writing a paper, we found it difficult to express our ideas in terms that would accurately transfer our thinking to the reader without leaving room for misunderstanding. Furthermore, the reduction to print of ideas as tenuous as the ones we expressed ordinarily lends an unwarranted rigidity which, in our case at least, is not intended. The following comments on points raised by Doctor Maxwell and Doctor Wise may possibly help to clarify our ideas.

Although in our view second-order, third-order, etc., shears can only be *generated* by movement on a primary shear, once the shear has been established and is in existence it must have an effect on continued deformation. We visualize the major elements of the shear pattern as having originated early in the history of the crust; subsequent deformation must have been controlled by the resulting segmentation.

Further, we recognize the possibility that east-west primary stresses may have existed rather than, or in addition to, north-south primary stresses, so that possibly in some areas, or at some time, the primary shears may have been parallel to the New Zealand and Texas directions. We simply do not know (p. 1232–1233).

The problem of determining which shear zones might actually be primary in the hierarchy of fault generations is a very difficult one, and its solution will probably not be attained for some years to come. In attempting to plot major shears known to, or suspected by, us on a globe, it appears that the only place on the crust of the earth where really extensive shears of the San Andreas direction are common is a belt along the western edge of North America and the western edge of the northern part of South America. If we regard the belt from the edge of the continental shelf off California to the eastern face of the Sierra Nevada as a mega-shear zone (which appears likely), we may be dealing with the only *primary* right-lateral shear zone in the earth's crust; all other elements of the San Andreas direction are possibly subsidiary in nature and subsequent in time of origin to this major parent shear zone. Such important earth features as the Oca, Bartlett, and Colombia faults (second-order) and the Texas lineament, Alpine fault, etc. (third-order) might well be the immediate first-generation offspring of the primary mega-shear zone mentioned above.

". . . the lower-order shears should be larger and more prominent. . . ." It is our experience that the lower the order of the shear, the more likely is it to be obscured by structural complexities, in an alluvium-covered valley or under water.

In regard to Alberding's commendable analysis of the tectonics in the area of Venezuela, we would venture the opinion that not all faults of the San Andreas direction, for example, need to be first-order shears. We pointed this out on page 1213 and again on page 1232.

We are not inclined to assign much significance to De Sitter's (and others') distinction between pure and simple (or rotational and nonrotational) shear, much as we admire his text otherwise. Our "primary stress" can be considered a resultant, or not—we have no way at the moment of seeing that far.

It is our belief that the roll and pitch of fault surfaces would necessarily result in local areas of lesser compressive stress, and even a local volume increase, which could accommodate the extensive brecciation and shattering associated with most wrench faults.

Vening Meinesz (1947) has calculated that present rotational forces would be sufficient to shear a crust of approximately half the strength of the present crust. It seems reasonable to us that in the early formative years of the crust it was much weaker than at present, and, hence, there could well have been a point in the earth's history at which the crust yielded to these forces. The continued existence of a sheared crust in this force field has finally resulted in the present-day complex tectonic aspect of the earth.

We are keenly aware of the sketchy nature of the information with which we have built

our concepts, but persist in our belief that the crust as a whole must be considered in any system of tectonics, and that there must be an understandable pattern of deformation when the earth is so considered. Rather than "a simple unidirectional system of global wrench-fault tectonics", our concept of the implications of our thinking is that of an ever-increasingly complicated pattern of deformation stemming from a relatively simple original stress system.

References Cited

Moody, J. D., and Hill, M. J., 1956, Wrench-fault tectonics: Geol. Soc. America Bull., v. 67, p. 1207–1246

Vening Meinesz, F. A., 1947, Shear patterns of the earth's crust: Am. Geophys. Union Trans., v. 28, p. 1–61

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WRENCH-FAULT TECTONICS: A REPLY

BY H. ALBERDING

Maxwell and Wise have made an interesting interpretation of the theory of wrench-fault tectonics of Moody and Hill (1956) and have raised some objections to the theory, especially as applied by Alberding (1957) to northern South America. They propose a different approach to the problem and suggest that major wrench faulting may take place by "simple" shear as defined by De Sitter (1956, p. 120–121).

(1) The proposal of Maxwell and Wise to interpret the orders of wrench faults of Moody and Hill as a hierarchy of fault types based on magnitude, with the greatest fault first, is not the only interpretation possible. Instead, it is suggested that the order of wrench faults of Moody and Hill is not of magnitude but of *sequence*, and the principal connection between fault orders is that the strain pattern caused by the release of stresses for one order controls the new distribution of stresses for the following order.

Moody and Hill (1956, p. 1211) mention that their system is different from McKinstry's (1953) whose "shears of the second order" are subsidiary to the "master" shear. For such a system "the available forces decrease rapidly and the system is not regenerative". Instead, Moody and Hill suggest a mechanism by which, following relief of primary stresses through folding, thrusting, and wrench faulting, a local stress reorientation occurs, and a secondary stress system is set up which finds partial relief by wrench movements along fractures already established by first-order wrench faulting. This process could proceed indefinitely, but only eight wrench directions result because of the values assigned to α, β and γ. Greatest stress release could occur along wrench faults of any order. As to the reason why the primary principal compressive stresses which initiate the system are meridional, Moody and Hill wisely refrain from hazarding a guess which could range from contraction of the earth coupled with changes in flattening at the poles to theories involving meridional convection currents.

(2) If it is true that the fault orders of Moody and Hill signify sequence instead of magnitude, then the objections raised by Maxwell and Wise do not apply to the writer's assignment of first order to the Tigre fault and second order to the major Oca fault in northern South America. Field evidence shows that the Tigre fault is earlier than the Oca (Rod, 1956, Fig. 2, p. 459). Note the drag of the Tigre fault and parallel structures along the south side of the Oca fault suggesting that the original extensions of these structures to the north have been displaced to the east by a right-lateral movement on the Oca fault. Geophysical work in the Gulf of Venezuela tends to confirm this.

(3) In regard to the suggestion of Maxwell and Wise that major wrench faulting takes place by "simple" shear as defined by De Sitter (1956, p. 120–121), the writer suggests that Billings (1954, Fig. 153, p. 179) gives a much better illustration of the relationship between faulting and the strain ellipsoid than does De Sitter (1956, Fig. 79, p. 121). As pointed out by Billings (1954, p. 103) "the strain gives us no direct evidence of the external forces that caused the deformation". De Sitter's (1956) Figure 79A could have been formed by couples just as well as by compression or tension, and his Figure 79B could have resulted from compression or tension just as well as from a couple. De Sitter (1956, p. 121) seems to realize this.

Further objection is made to the obtuse angle of 120° shown by De Sitter in Figure 79A between the shear planes bisected by the least-strain axis of the strain ellipsoid (the axis of greatest principal stress). Assuming that the rock was nonplastic and that no rotation occurred following initial stress release, the angle between the shear planes bisected by the greatest principal stress should be acute (Anderson, 1951, p. 62; Billings, 1954, p. 103, 179; McKinstry, 1955, p. 178; Moody and Hill, 1956, p. 1209–1210).

In any event, the supposition made by Maxwell and Wise that the mere presence of a single large wrench fault suggests *ipso facto* that "simple" shear by a couple was involved is suspect. This is especially true when applied to the Oca-El Pilar faults of northern South America that have a complementary wrench fault in the Great Fault of Colombia (the

"Gran Falla" of Raasveldt, 1956, Fig. 5, p. 25 or Santa Marta fault of Young *et al.*, 1956, Fig. 9, p. 24). A secondary principal compressive stress in a N. 60° W. direction was assumed by the writer to account for movements on these faults, once their directions were established by the strain pattern formed by displacements on first-order faults of Great Glen direction, as represented by the Tigre fault (Alberding, 1957, Fig. 2, p. 787).

(4) In regard to the statements of Maxwell and Wise that "in pure shear—the 'force' normal to the shear plane bears a direct relation to the sine of the angle of shear", and, when the angle of shear is equal to 30°, the normal "force" would be one-half the primary compressive "force", the writer refers to Hafner (1951, p. 376–377) who cautions against confusing "forces" with "stresses" and solving stress problems by simple trigonometry or vector addition. The pressure or tension normal to a shear plane is a *stress* that exists regardless of whether the external forces which set up the stress system were compressional, tensional, or a couple. The normal stress is related to the stress ellipsoid; it ranges from a maximum to a minimum (Billings, 1954, p. 100–101) but never becomes zero as it would, according to Maxwell and Wise, when the shear angle is zero. Anderson (1951, p. 9–11) shows the normal stress to be equal to $P \sin^2\theta + R \cos^2\theta$ where P = greatest stress, and R = least stress and θ = the angle of shear (between the greatest stress and the shear plane). If $\theta = 0$, $\cos^2\theta = 1$, and the normal stress would not be zero.

(5) Anderson (1951, p. 19–20) discusses brecciation along faults which depends on the relationship of the normal pressure across the fault plane to the standard or vertical pressure for a given depth. If the normal pressure across the fault is less than the vertical pressure, as is the usual case along wrench faults, then brecciation can occur. Along thrust faults the normal pressure across the fault is always greater than the vertical pressure, and along normal faults it is always less.

In conclusion, it appears that despite the objections raised by Maxwell and Wise to the theory of wrench-fault tectonics of Moody and Hill, especially as applied by Alberding to northern South America, the theory still seems preferable to any proposed mechanism of "simple" shear. The theory of Moody and Hill has such points in its favor as explaining not only the orderly sequence of wrench faulting in northern South America, but of accounting for the N. 75° E. strike of the folds and thrusts of the Cordillera de la Costa and Serrania del Interior by primary principal compressive stresses acting at right angles (α = 345°). This in itself would make the theory of Moody and Hill preferable to any which tried to explain these folds and thrusts, which cover a north-south width of 120 km at the longitude of Caracas, as secondary drag effects from movements along an east-west fault (Rod, 1956, Fig. 11, p. 473).

As to where the second-order and first-order wrenches are in New Zealand is a matter to be delved into by geologists familiar with the area, but the region to the north and northwest of the islands appears to offer possibilities.

References Cited

Alberding, H., 1957, Application of principles of wrench-fault tectonics of Moody and Hill to northern South America: Geol. Soc. America Bull., v. 68, p. 785–790

Anderson, E. M., 1951, The dynamics of faulting: 2nd ed., Edinburgh, Oliver and Boyd, 206 p.

Billings, M. P., 1954, Structural geology: 2nd ed., Englewood Cliffs, N. J., Prentice-Hall, 514 p.

De Sitter, L. U., Structural geology: N. Y., McGraw-Hill, 552 p.

Hafner, W., 1951, Stress distributions and faulting: Geol. Soc. America Bull., v. 62, p. 373–398

McKinstry, H. E., 1953, Shears of the second order: Am. Jour. Sci., v. 251, p. 401–414

—— 1955, Structure of hydrothermal ore deposits: Fiftieth Anniv. Vol., Econ. Geology, Part I, p. 170–225

Moody, J. D., and Hill, M. J., 1956, Wrench-fault tectonics: Geol. Soc. America Bull., v. 67, p. 1207–1246

Raasveldt, H. C., 1956, Fallas de rumbo en el Nordeste de Colombia: Revista del Petroleo (Bogotá),. v VII, no. 64, p. 19–26

Rod, E., 1956, Strike-slip faults of northern Venezuela: Am. Assoc. Petroleum Geologists Bull., v. 40, p. 457–476

Young, G. A., Bellizzia, A., Renz, H. H., Johnson, F. W., Robie, R. H., Mas Vall, J., 1956, Geología de las cuencas sedimentarias de Venezuela y de sus campos petroliferos: Bol. de Geol. (Venezuela), Pub. Esp. no. 2, 140 p.

SIGNAL OIL AND GAS OF VENZEUELA, APARTADO DEL ESTE 5484, CARACAS, VENEZUELA

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APPLICATION OF PRINCIPLES OF WRENCH-FAULT TECTONICS OF MOODY AND HILL TO NORTHERN SOUTH AMERICA

BY EMILE ROD

H. Alberding (1957, p. 785) writes:

"The purpose of this paper is twofold: (1) to point out the ideal way in which the principles of wrench-fault tectonics expressed by Moody and Hill (1956, p. 1207–1246) apply to northern South America (northern Venezuela, Trinidad and northeastern Colombia); and (2) to present evidence which suggests movements of land masses along the strike-slip (wrench) faults along the southern edge of the Caribbean in northern South America comparable and complementary to those movements which occurred, according to Hess and Maxwell (1953, Fig. 2, p. 4), along strike-slip (wrench) faults in the Greater Antilles and Central America in the northern part of the Caribbean."

Alberding's ideas are illustrated in two figures and one plate. Figure 2 (p. 787) shows Plan of Wrench-Fault System developed in Northern South America. In the text we note that a primary compressive stress oriented N. 15° W. has been assumed (p. 789).

"Figure 2 shows a plan of strike-slip (wrench) fault system developed in northern South America according to the principles of Moody and Hill (1956, Fig. 5, p. 1213). The primary compressive stress is assumed to have been exerted in a N. 15°W. direction ($\alpha = 345°$) so that the various fault directions should be oriented correctly (Moody and Hill, Fig. 20, p. 1234)."

At the bottom of the same column Alberding writes as if the direction of the primary stress had been well established and is now a fact. He seemingly firmly believes that the direction of primary stress and the fault system, as shown on Figure 2, are the only and final solution. Hence he concludes that:

"In northern South America, the Great Glen and San Andreas first-order shear directions (Fig. 2, *I*) are represented by the Tigre fault (a left-lateral Great Glen direction wrench) and the San Francisco and Los Bajos faults (right-lateral San Andreas direction wrenches)."

That the Great Glen and San Andreas shear directions should be of first order is another statement which seems to be based on a misunderstanding as to the significance of the "eight major wrench directions" (1956, p. 1213) of Moody and Hill. Strike-slip faults parallel to any of those eight directions can be of any order. Depending on the regional tectonic system a fault oriented approximately N.15°E.–N.35°E. (Moody and Hill, 1956, p. 1228) may be assigned to the "Great Glen direction". In area *A* it can represent a master fault of first order, but in area *B* it might be of second or third order.

Thus, as a result of assumptions, Alberding now raises typical second-order shears, such as the Tigre, San Francisco, and Los Bajos faults, to first-order strike-slip faults and degrades the master faults to second-order shears, and this in spite of the special emphasis and careful discussion which is given to "second-order effects" in the paper by Moody and Hill.

The hypothesis of Moody and Hill is actually based on the recognition of master faults and the second-order shears caused by movements of the crustal blocks along master faults. Which is the master fault; this should be the first question when starting to analyze the regional fault pattern.

The Urica, San Francisco, and Los Bajos faults do not cut the El Pilar fault (Rod, 1956a; 1956b). They start a few kilometers south of the El Pilar fault as right-lateral, southeast-trending strike-slip faults, swing gradually in a more easterly direction, and turn finally into the east-northeast striking frontal thrust zones. Kugler's map and cross section of Trinidad (1956, Pl. 1) show this relationship very clearly.

A textbook example of a scoop-shaped second-order strike-slip fault turning into a thrust is the Urica fault forming the west-southwest boundary of the Serrania del Interior of Eastern Venezuela (Rod, 1956a, Figs. 1, 9; 1956b, Fig. 1). In the Tacat area the direct continuation of the Urica fault into the Pirital thrust has been proved by seismic surveys and several deep tests. For the simple reason that the permission of at least four oil companies would be needed to publish the pertinent facts, no detailed description was presented. Alberding (1957, p. 785), however states: "the writer does not accept the interpretation that numerous subsidiary strike-slip

faults turn into thrusts (Rod, 1956, Fig. 1, p. 458)".

Several textbooks on tectonics mention the close association of strike-slip faults with thrusts. On page 1221, Moody and Hill writer postulates: Rod shows an east-west strike-slip movement along the El Pilar fault (a second-order right-lateral wrench) as causing the development of a first-order right-lateral wrench (the San Francisco fault of San Andreas direction of Moody and Hill nomenclature)."

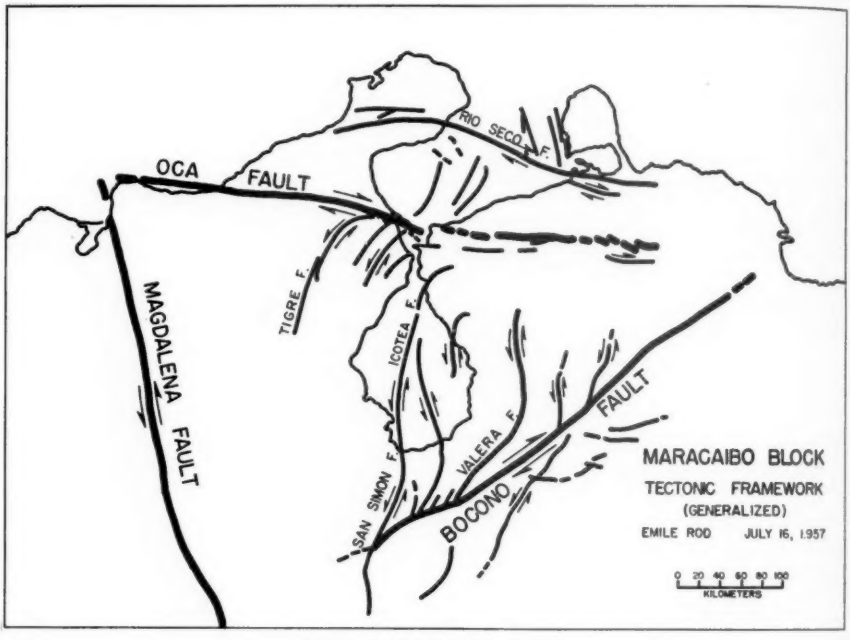

FIGURE 1.—MARACAIBO BLOCK
Fault system after maps published by E. Rod, H. G. Raasveldt, O. Renz, and H. Alberding

(1956) describe a few examples from California.

Bucher (1955) emphasizes such a relationship. Referring to the Mojave block northeast of Barstow (from Hewett's description) he writes (1955, p. 362):

"At least the northern half of the triangle's base is a low-angle thrust fault which, on the northeast slope of the Arawatz Mountains, joins the Garlock fault in such a way as to leave no room for doubt that it records the differential eastward movement of the wedge block," and on the Lake Basin zone (p. 363): "As in the Mojave desert, here too overthrusts, trending at right angles to the great fault zone, swing into it abruptly taking up the effect of the differential movement."

To be consistent with requirements of his plan (Fig. 2), Alberding had to reverse the genetic relationship of the faults:

"The development of shear planes as shown by Rod (1956, Fig. 11, p. 473) is just opposite from what the

Although Moody and Hill (1956, p. 1232, 1233) favor a north-south orientation for the primary principal compressional stresses they do not leave the least doubt that this is just a hypothetical explanation.

A northwest-southeast orientation of the primary compressional stresses in the basement blocks south of the Caribbean Sea, and a northeast-southwest orientation north of the Caribbean could best explain the observed master faults bounding the Caribbean Sea to the north and to the south. Such a stress orientation would suggest two approximately east-west couples (counter clockwise to the north and clockwise to the south of the Caribbean axis) as external forces, one force, common to both couples, active in the long axis of the Caribbean from west to east.

Instead of considering the movement of basement blocks along their boundary faults

(Cloos' geofractures, or major strike-slip faults) Alberding (1957, p. 788) jogs around minor morphologic units such as the Guajira Peninsula.

The Maracaibo block[1] is a polygonal segment of the crust bounded by the Oca, Magdalena,[2] and Boconó strike-slip faults (Fig. 1). A crustal block may change in shape but is displaced as a unit along its boundary faults. A carving knife would be needed to move the Guajira Peninsula from position (1) (Alberding 1957, p. 788, and Fig. 1) to position (1a).

Thus Alberding's "ideal way" in which the "principles" of wrench-fault tectonics expressed by Moody and Hill (1956, p. 1207–1246) apply to northern South America consists in presenting hypothetical explanations and confusing cause and effect.

In vain we look in Moody and Hill's paper for anything which might be interpreted as "principles of wrench-fault tectonics". The authors present a thesis and develop a hypothesis which still has to be proved or disproved.

[1] Dufour's Maracaibo block includes only the Maracaibo Basin and is bounded on the southeast by the Mérida Andes and on the west-northwest by the Perijá Mountains.

[2] For Raasveldt's (1956, p. 23) "Gran Falla de Colombia" or "Gran Falla" the name "Magdalena fault" is proposed after the ancestral valley of Rio Magdalena which has been offset by the fault. A geographical name restricted to a smaller area than "Colombia" would seem preferable. The northern half of the Magdalena fault was shown in 1933 on De Cizancourt's maps (1933, Figs. 1, 3). Only the northernmost quarter of Raasveldt's Gran Falla de Colombia (here Magdalena fault) coincides with the Santa Marta fault of Young.

North of El Banco the two faults diverge to such a degree that near their southern ends the east-west separation of the fault traces measures almost 120 km.

References Cited

Alberding, H., 1957, Application of principles of wrench-fault tectonics of Moody and Hill to northern South America: Geol. Soc. America Bull., v. 68, p. 785–790

Bucher, W. H., 1955, Deformation in orogenic belts, p. 343–368 *in* Poldervaart, Arie, *Editor*, Crust of the earth: Geol. Soc. America Special Paper 62, 762 p.

Cizancourt, Henry de, 1933, Tectonic structure of northern Andes in Colombia and Venezuela: Am. Assoc. Petroleum Geologists Bull., v. 17, p. 211–228

Dufour, J., 1955, Some oil-geological characteristics of Venezuela: 4th World Petroleum Congress Proc., Rome

Kugler, H. G., 1956, Trinidad, p. 353–365 *in* Jenks, W. F., *Editor*: Handbook of South American Geology: Geol. Soc. America Mem. 65, 378 p.

Moody, J. D., and Hill, M. J., 1956, Wrench-fault tectonics: Geol. Soc. America Bull., v. 67, p. 1207–1246

Raasveldt, H. C., 1956, Fallas de rumbo en el Nordeste de Colombia: Revista del Petroleo (Bogotá), v. VII, no. 64, p. 19–26

Rod, Emile, 1956a, Strike-slip faults of northern Venezuela: Am. Assoc. Petroleum Geologists, Bull., v. 40, p. 457–476

—— 1956b, Earthquakes of Venezuela related to strike-slip faults?: Am. Assoc. Petroleum Geologists Bull. ,v. 40, p. 2509–2512

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WRENCH-FAULT TECTONICS: A REPLY

BY H. ALBERDING

Emile Rod has made a number of comments relating to the writer's paper on *Application of principles of wrench-fault tectonics of Moody and Hill to northern South America* and has suggested that two primary compressional stresses, acting in northeast–southwest and northwest–southeast directions, could best explain the observed master faults bounding the Caribbean Sea to the north and to the south. Rod further suggests that such a stress orientation could have resulted from two approximately east-west couples (counterclockwise to the north and clockwise to the south of the Caribbean axis) and that one external force, common to both couples, was active along the long axis of the Caribbean from west to east.

In reply to what Rod has written, the writer would like to make the following remarks:

Rod objects to the word "principles" as applied by the writer to Moody and Hill's theory of "Wrench-fault tectonics", but even a theory can have certain principles upon which it is based.

One of the most fundamental mechanical principles used by Moody and Hill in their theory is that, if an outside external compressional force had been exerted from west to east along the long axis of the Caribbean, a *single* primary principal compressive stress would have resulted along an east-west axis and *not* two primary compressive stresses along northeast–southwest and northwest–southeast axes as suggested by Rod. Apparently, in deriving these two stresses from a single external force, Rod had in mind the resolution of forces by which a single force can be resolved into two components by constructing a parallelogram (Billings, 1954, p. 10). This, however, involves forces throughout, not one external force and two stresses. Hafner (1951, p. 376–377) warns against expressing stresses as vector quantities and deriving them by vector addition such as can be done with forces in the force parallelogram.

Another basic "principle" for the rupture of brittle substances used by Moody and Hill, but ignored by Rod in his above analysis, is that, if an external compressive force had exerted itself across the long axis of the Caribbean from west to east, and if an east-west-trending primary principal compressive stress had resulted from this external force, any first-order wrench faulting would have occurred at an angle of 30° (β) to the principal compressive stress axis. Assuming that a complementary set of shears developed, strike-slip movements along these first-order right- and left-lateral wrenches could have produced fractures along which, following a local reorientation of stresses, second-order right- and left-lateral strike-slip movements could have occurred. Two of these second-order wrench directions would follow the Bartlett and Oca directions of Moody and Hill's terminology.

Moody and Hill (1956, p. 1232) state "that the present evidence indicates that the primary principal compressional stress direction has not varied much more than 20° to the west or east of north . . .". They further say: "However, a symmetrical distribution of shear directions would result from primary compressional stresses oriented east-west which might exist in the earth's crust".

However, regardless of the direction of the primary principal compressional stress axis, the Oca and Bartlett directions of Moody and Hill, if considered as complementary, which they certainly are in the Caribbean, would always be second order or lower order.[1] This is obvious because these two fault directions, although one is right lateral and the other is left lateral, are almost parallel to one another and thus could not be a complementary set of shears with a 60 degree angle between them (both 30° to the principal compressive stress axis).

Rod repeats what Moody and Hill (1956, p. 1228) had to say about giving the various fault directions a certain leeway in their strike (*e.g.*, San Andreas direction, N. 30° W. to N. 45° W.; Great Glen direction N. 15° E. to N. 35° E., *etc.*). However, this does not mean that just because a Great Glen direction fault can vary 20° in strike from N. 15° E. to N. 35° E. it could change, as stated by Rod, from a first-order fault in area "A" to a second- or

[1] Much confusion exists in this discussion because Maxwell and Wise label first-order wrenches low order and all other wrenches progressively higher order, whereas Moody and Hill consider first-order wrenches high order and all others progressively lower order.

third-order fault in area "B". The order of any fault is determined by what happened before and after it, not its possible variation in strike.

As to Rod's and Moody and Hill's (personal correspondence) criticism of the writer's designation of the Tigre fault as a first-order fault and the Oca as a second-order fault, the writer can only say that when a fault, such as the Tigre, is cut off and dragged and to all appearances displaced to the east by another fault (the Oca), such a fault certainly was in existence *before* the fault which cut, dragged, and displaced it. Rod calls the Tigre fault second order and, therefore, later than the Oca fault which he calls first order. Moody and Hill call the Oca fault second order and the Tigre fault third order or lower order.

The writer is fully aware that not all faults of San Andreas direction need be first-order wrenches. However, it seems much more plausible to associate the San Francisco and Los Bajos faults of San Andreas direction in northeastern Venezuela and Trinidad with the primary principal compressive stresses which must have existed in a N. 15° W. (α = 345°) direction at right angles to the N. 75° E.-striking folds and thrusts of the Cordillera de la Costa and Serranía del Interior and the ranges of Trinidad. Rod (1956, Fig. 11, p. 473) interprets the wide belt of folds and thrusts in northeastern Venezuela, as well as the major strike-slip faults of San Andreas direction in this region, as secondary drag effects from movements on the El Pilar strike-slip fault. It appears to the writer that Rod overemphasizes the effects of strike-slip faults and attempts to associate practically every structure in northern Venezuela and Trinidad with strike-slip movements.

This likewise applies to Rod's Figure 1 in this discussion; again the writer believes that Rod is overemphasizing strike-slip faults. Many of the strike-slip faults which Rod shows as offshoots of the Bocono fault are known to be normal faults and even thrusts (compare the writer's Pl. 1, 1957, with Rod's Fig. 1 in this discussion).

Rod's extension, on his Figure 1 in this discussion, of the Rio Seco fault across the Gulf of Venezuela and into the Cocinas fault of O. Renz (1956, Fig. 3) is too simplified an interpretation. Also, the name Santa Marta fault (Young *et al.*, 1956, Fig. 9, p. 24) should have precedence over Rod's proposed name Magdalena fault for the "Gran Falla" of Raasveldt (1956, Fig. 5, p. 25), which last name admittedly should be dropped.

Referring to the writer's representation in Figure 1 (Alberding, 1957, p. 786) of the Guajira Peninsula having moved (relatively) from position 1 to 1a, and which led to Rod's remark that a carving knife would be needed to perform this movement; the writer had no intention of implying that the rocks of the Guajira Peninsula were in position 1 detached from the remainder of what is now the northern coast of South America. The intention was to represent the strike-slip movement along the Great Fault of Colombia (the Santa Marta fault of Young *et al.*) by showing the relative displacement of the rocks of the Guajira Peninsula. To have shown more than the peninsula would have cluttered up Figure 1. Needless to say, at the time of the movement, the Guajira Peninsula did not have its present geographic outline.

In conclusion, the writer is certainly aware that Moody and Hill's article on *Wrench-fault tectonics* presented a theory; the writer merely attempted to apply the theory to northern South America and the Caribbean.

However, even though Moody and Hill's "Wrench-fault tectonics" is a theory, it is based on certain fundamental mechanical principles which are sound for the rupture of brittle substances. Theories based on plastic deformation, such as suggested by Bijlaard (1936) and Vening Meinesz (1947, p. 6-28) and experimented with by E. Cloos (1955, p. 241–256), would involve a different approach to the problem.

References Cited

Alberding, H., 1957, Application of principles of wrench-fault tectonics of Moody and Hill to northern South America: Geol. Soc. America Bull., v. 68, p. 785–790

Bijlaard, P. P., 1936, Theorie des deformations plastiques et locales par rapport aux anomalies negatives de la gravitation, aux fosses oceaniques, aux geosynclinaux, au volcanisme, a l'orogenie et a la geologie de l'ocean pacifique occidental: Assoc. Geodesie, Cong. d'Edinbourg

Billings, M. P., 1954, Structural geology: 2d ed., New Jersey, Prentice-Hall, 514 p.

Cloos, E., 1955, Experimental analysis of fracture patterns: Geol. Soc. America Bull., v. 66, p. 241–256

Hafner, W., 1951, Stress distributions and faulting: Geol. Soc. America Bull., v. 62, p. 373–398

Moody, J. D., and Hill, M. J., 1956, Wrench-fault tectonics: Geol. Soc. America Bull., v. 67, p. 1207–1246

Raasveldt, H. C., 1956, Fallas de rumbo en el Nordeste de Colombia: Revista del Petroleo (Bogotá),. v VII, no. 64, p. 19–26

Renz, O., 1956, Cretaceous in western Venezuela

and the Guajira (Colombia): Paper presented at 20th Internat. Geol. Cong., Mexico City, September

Rod, E., 1956, Strike-slip faults of northern Venezuela: Am. Assoc. Petroleum Geologists Bull., v. 40, p. 457–476

Vening Meinesz, F. A., 1947, Shear patterns of the earth's crust: Am. Geophys. Union Trans., v. 28, p. 1–61

Young, G. A., Bellizzia, A., Renz, H. H., Johnson, F. W., Robie, R. H., Mas Vall, J., 1956, Geología de las cuencas sedimentarias de Venezuela y de sus campos petrolíferos: Bol. de Geol. (Venezuela), Pub. Esp. Nº 2, 140 p.

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TERMINOLOGY OF POST-VALDERS TIME

BY WILLIAM S. COOPER

The most recent phase of Pleistocene history, essentially the period from culmination of the latest major advance of Wisconsin glaciation (here taken to be the advance which in Wisconsin and neighboring regions is designated Valders[1]) to the present, has acquired a set of terms—one cannot speak of a consistent, generally accepted nomenclature—that have appeared in haphazard fashion and are confusing to a high degree. The inappropriateness of some has been pointed out by several authors, recently by Judson (1953, p. 59) and by Deevey and Flint (1957, p. 182). Terms that have attained a degree of currency die hard. Objections previously made will gain by reinforcement; attention should be called to those that so far seem to have escaped criticism. The first objective in this paper is a critical review of the terms that have found their way into common use. The second is submission of a small-scale terminology fitted particularly to the period concerned, with the hope that it will be found more useful than the present chaotic miscellany.

I agree with Flint (1957, p. 281–284) that *Tertiary Period* and *Quaternary Period* should be dropped from stratigraphic nomenclature and that the Pleistocene be considered an epoch of the "initial period" (given no name by Flint) of the Cenozoic era, co-ordinate with the Pliocene and earlier epochs, and extending to the present day. This viewpoint automatically eliminates the term *Post-Pleistocene*. The term *Recent*, denoting a formal unit co-ordinate with Pleistocene, is also ruled out. *Postglacial* "can properly be used only in a local sense and in an informal way" (Flint, 1957, p. 283). With respect to validity of the term *Post-Wisconsin*, in order to be quite sure, a waiting period of some thousands of years is prescribed. All the terms commonly applied to the period under consideration as a whole have thus been eliminated.

As to the general climatic sequence during this period there seems to be practically universal agreement. Three phases or subunits are recognized, which are essentially those formulated descriptively, but not named, by Von Post (1931, p. 53): a period of rising temperatures leading to one during which temperatures were somewhat higher than at present; this followed by a time characterized by a moderate decrease in temperature, extending essentially to the present day. It is commonly assumed that the middle period, at least part of the time and in many regions, was the driest of the three.

There have of course been fluctuations in climate, both in temperature and precipitation, superposed upon the broadly generalized sequence. In North America the Cochrane episode, near the end of the first subunit, seems to have been one of these; the "multiple glaciation since the ice age" (Harrison, 1956, p. 181; *see also* Karlstrom, 1955, p. 1581; 1956, p. 310) fits with the third. Fluctuations of considerable magnitude doubtless occurred even during the warm middle period. The Blytt-Sernander scheme, which persists in European and to some extent in American literature in spite of its awkward terminology, subdivides the warm middle subunit into a wetter phase (Atlantic) preceded and followed by a drier (Boreal and Sub-Boreal respectively). "Recurrence horizons" are assumed to indicate fluctuations of moisture during and somewhat before the latest of the three subunits.

We next review the terms that have been applied to the three subunits. The first subunit, characterized by rising temperatures, has been neglected by the term makers, in spite of the magnitude of the phenomena associated with it. A single proposal, *Anathermal*, covers a period partially co-extensive with the subunit as here defined and is reserved for later discussion.

The middle subunit, characterized by maximum warmth, has suffered from no such neglect. Of the several terms that have attained more or less currency, *Climatic Optimum* is the least acceptable. This anthropocentric expression connotes, in the words of Ahlmann (1953, p. 37), "a milder regime, more favorable to mankind." Do we know how *Homo sapiens* of

[1] It is realized fully that the sequence of events from Cary to Valders is as yet far from clear; and also that Antevs (1953, p. 203), Karlstrom (1956, p. 326), and Just (1957) treat the Cochrane episode rather than the Valders as the latest major advance. Recent evidence from various fields of research seems to point clearly to the Valders as the turning point in late climatic history (Broecker, 1957).

that period felt about it; and what does it matter how we feel? For other animals and for plants which were adjusted to relatively low temperatures, evicted by the newcomers, it was *pessimum*. In spite of occasional protests the phrase persists; it appears in at least eight important papers by American authors issued between 1953 and 1957. In some of these, enclosure in quotation marks suggests hesitation as to its suitability. It finds unquestioned acceptance in the very recent comprehensive work by Charlesworth (1957, p. 1482–1500). The corollary expressions "climatic improvement" and "climatic amelioration" (rising temperatures) and "climatic deterioration", "degeneration", "worsening" (colder and wetter) must obviously stand or fall with "Climatic Optimum." Certain authors who have abandoned "Climatic Optimum" still cling to its corollaries.

Flint and Deevey (1951, p. 275) expressed preference—which they have since revoked (Deevey and Flint, 1957, p. 182)—for *Thermal Maximum* over *Climatic Optimum*. The phrase is objective and descriptive; standing by itself, however, with no companion terms for the periods before and after, it seems a bit lonely. *Xerothermic* is perhaps as old an appellation as any and appears frequently even to the present. The connotation of dryness as an essential element renders it questionable. Also, it has been treated by some authors as distinct from and later than *Climatic Optimum* (P. W. Smith, 1957, p. 206) or as a constituent subunit of the latter, equivalent to Sub-Boreal of the Blytt-Sernander scheme (Charlesworth, 1957, p. 1482–1500). Like *Thermal Maximum* it stands without companion terms, as does *Megathermal*, proposed by Judson (1953, p. 59). *Altithermal* of Antevs and *Hypsithermal* of Deevey and Flint will be considered later.

The period of moderate cooling, with limited glacier expansion and, hypothetically at least, a small eustatic lowering of sea level, covering the last few millennia, has acquired a designation that has found wide use: the *little ice age*. This was introduced into the literature by Matthes (1939, p. 520): "It may well be said that we are living in an epoch of renewed but moderate glaciation—a 'little ice age' that already has lasted about 4,000 years." A year later Matthes published a more extensive treatment, and incidentally revealed the source of his designation (Matthes, 1940, p. 398): "There appears to be warrant for the assertion that the present age is witnessing a mild recrudescence of glacial conditions—that it is, as a clever journalist has suggested, a separate 'little ice age.'" The term "caught on" and has been used in many publications from the time of its first appearance to the immediate present. Usually it has been surrounded by quotation marks, indicating that authors have thought of it as an informal stop-gap sort of expression; surely Matthes must so have considered it. Several, however, have formalized it by capitalization and omission of quotation marks.

The expression *little ice age* is misleading because, even with the qualifying "little", it exaggerates the magnitude of glacier expansion during the period so designated. The expression "ice age" has been almost universally applied to those few known major periods in geologic history marked by extensive, complex glaciation, and in particular to the Pleistocene (for example, Coleman, 1929; Daly, 1934; Woldstedt, 1954); its application to this minor episode is to be deplored. One has only to recall the extremely slight lowering of sea level during the "little ice age" in comparison with the impressive rise associated with melting away of the temperate continental ice sheets to realize that this "ice age" was less than little; it was insignificant. The expression clashes, too, with the generally recognized use of "age" for a definite geologic-time unit (American Commission on Stratigraphic Nomenclature, 1952, p. 1636).

Matthes inadvertently added to the confusion by once using the expression "lesser ice age" (Matthes, 1942, p. 207). This has been taken by Karlstrom (1956, p. 310), and perhaps by others, to refer to an event distinct from and later than the "little ice age." Careful reading of Matthes' statement in context will show clearly that no separation was intended: "Accordingly he [Kinzl, 1932, p. 269] feels justified in declaring that the glacier advances of the last 300 years are the greatest that have occurred since the Pleistocene ice age. Those 300 years therefore comprise really a separate epoch of glacier expansion, a lesser ice age, that was preceded by a warm period of considerable duration. The occurrence of that warm period, in the middle of the Post-Pleistocene interval. . . ."

Two other terms have been applied to the third subunit, marked by moderate lowering of temperature: *Katathermal* by Chiarugi (1936, p. 3) and *Medithermal* by Antevs (1948, p. 176). Charlesworth's heading in discussing the period is "Sub-atlantic degeneration" (Charlesworth, 1957, p. 1495).

Having eliminated (at least to his own satisfaction) most of the terms in common use for the latest period in Pleistocene history, it is incumbent upon the author to attempt construction of a satisfactory set of substitutes. In this situation the most useful course seems to be to set aside the strict canons of geologic nomenclature and devise a set of provisional terms that will serve our present purpose and that may be superseded whenever in the future

TABLE 1.—PROVISIONAL SYSTEM OF TIME UNITS PROPOSED IN THIS PAPER, ADJUSTED TO DANISH POLLEN-ZONE AND BLYTT-SERNANDER SEQUENCES

Upper and lower limits of Hypsithermal according to Deevey and Flint (1957, Table 1). Date for Valders advance follows Flint (1957, Table 23-B, p. 395).

POLLEN ZONES	BLYTT-SERNANDER	C^{14} YEARS B. P.	PROVISIONAL TIME UNITS	
		1000		
IX	SUB-ATLANTIC		HYPOTHERMAL	
		2000		
			2600	
		3000		
VIII	SUB-BOREAL			
		4000		
		5000		
VII		6000		
	ATLANTIC		HYPSITHERMAL	NEOTHERMAL
		7000		
VI		8000		
V	BOREAL	9000	9500	
IV	PRE-BOREAL	10000	ANATHERMAL	
III	YOUNGER DRYAS	11000	10700	
II	ALLERÖD		VALDERS-FENNOSCANDIAN	

The results of this effort are presented graphically in Table 1. Before entering on discussion it is necessary to establish certain premises.

The shortness of the period considered (about 11,000 C^{14} years) and its closeness to us make it difficult to utilize the terminology standard for geologic periods that are longer and farther from us. Our descendants 15,000 or 20,000 years from now (if any) will put us in our proper time unit, with approximate dates setting off ours from the units preceding and following. The difficulty for us of the present is that we do not know whether we are in a waning phase of a glacial age, or part way along in an interglacial, or on our final way out from the grand period of Pleistocene glaciation.[2]

a longer view becomes possible. There will always be a period, as decades and centuries pass, when provisional terminology will be useful. For future scientists, as for us, it will be helpful to devise provisional terms to cover the events close to them in time.

As foundation for the new terminology it is possible to utilize the well-known set of terms proposed by Antevs (1948, p. 176) to cover the period variously referred to as Postglacial, Post-Pleistocene, and Recent. The major unit in this scheme is the *Neothermal age*, which "should have the same rank as the Wisconsin Glacial and the Sangamon Interglacial (Sanga-

[2] Certain authors are willing to venture a guess. Karlstrom (1956, p. 327) suggests that the Altithermal be considered "to fall near the culmination of an interglacial period which separates the Wisconsin glacial stage from a still unnamed glacial stage that may attain its maximum extent within the next 20,000 years." Charlesworth (1957, p. 1535) offers a similar suggestion.

mon Thermal)." It comprises three subunits, the *Anathermal*, "at first as today but growing warmer", the *Altithermal*, "distinctly warmer than at present", and the *Medithermal*, "moderately warm", ages. In a later paper Antevs (1953, p. 195) adds the term *Deglacial* age, which apparently covers in North America the time from the greatest extension of Wisconsin glaciation to the beginning of recession from the Cochrane moraines. Antevs applies the term "age" without regard to rank; his Deglacial age overlaps a considerable part of the Wisconsin age as commonly understood. In utilizing his scheme substitutions and a reinterpretation are made. There is no implied acceptance of his absolute chronology.

Neothermal is an appropriate designation for the major time unit treated in this paper. It is here used as a provisional term for the time since culmination of the latest major advance of Wisconsin glaciation, assumed to be the Valders in central North America, correlated with the Fennoscandian in northwestern Europe. It is not, as in Antevs' use, co-ordinate with the Sangamon and the other interglacials.

Past climatic conditions must be inferred from material evidence, and of this there is here no lack: there are drift sheets, bog deposits, deep-sea sediments.

For the first subunit, marked by rising temperature, the term *Anathermal* is satisfactory;[3] its beginning, however, is set at the end of the Valders advance rather than at the end of the Cochrane.

Antevs' term Altithermal for the middle subunit, with temperatures somewhat higher than at present, is, as Deevey and Flint point out (1957, p. 182), an "etymologic hybrid." These authors in the same paper, rejecting Climatic Optimum, Thermal Maximum, Xerothermic, Altithermal, and Megathermal, propose *Hypsithermal*, adapted from Chiarugi's *ipsotermico* (1936, p. 3).[4]

Since the term has been incorporated into Flint's recent textbook (1957, p. 396) and has been utilized in several very recent papers, it seems wise to accept it. Deevey and Flint (1957, p. 182) define the Hypsithermal as "the time represented by four pollen zones, V through VIII in the Danish system", which are equivalent to the Boreal, Atlantic, and Sub-Boreal of the Blytt-Sernander sequence and thus to Von Post's "post-glacial warm period" as defined in his latest statement (1946, p. 194). According to Deevey and Flint (1957, p. 184) the Hypsithermal is dated in C^{14} years "approximately from 7000 B.C. to 600 B.C." (In Table 1 the lower limit is apparently 7500 B.C.) Flint in his textbook (1957, p. 377) gives different bounding dates: 7500–4000 B.P. (5500–2000 B.C.). The limits given by Deevey and Flint seem to fit most conveniently the stratigraphic sequence.

These dates automatically fix also the end of the Anathermal and the beginning of the third, final subunit. The approximate 2000 (or 1500) years thus allowed to the Anathermal are altogether too few for disappearance of the temperate continental ice sheets; they must have lasted well into Hypsithermal time. This assumption is supported by data cited by Flint (1957, p. 262) indicating that 1000–2000 years after the beginning of Hypsithermal time as set by Deevy and Flint sea level was still at least 36 m lower than now. I cannot agree with Flint (1957, p. 470) and Deevey and Flint (1957, p. 184) that the sequence of pollen zones assigned to the Hypsithermal may on present knowledge be considered a substage; only our rather distant descendants can settle that point.

Antevs' term *Medithermal* for the final subunit is another "etymologic hybrid." Deevey and Flint apparently think that the period needs no name. The time left for it is admittedly comparatively short—2600 years as against 6400–6900 for the Hypsithermal—and only a single pollen zone (IX) and a single Blytt-Sernander unit (Sub-Atlantic) are involved. It is true, further, that deep-sea sediments have so far apparently yielded no evidence for a period of lowered temperature following the Hypsithermal, and the testimony of lowered sea level is equivocal. Lack of convincing evidence here is not strange in view of the small magnitude of the climatic change. There is, however, material evidence on the continents for a period distinct climatically from the Hypsithermal: in till laid down by advancing glaciers (exposed by very recent recession) and particularly in pollen stratigraphy. If this were not so, there would be no reason for setting up a Hypsithermal interval. The subunit, even though comparatively insignificant with respect to magnitude of events and length of geologic time, needs a name because it is so close to us; repeated use of the unsatisfactory "little ice age" is proof of it. The appellation *Hypothermal* is here proposed.

[3] Emiliani (1955, p. 547) employs this term in a completely different sense: *anathermal* and *catathermal* are used to define the phases of rising and declining temperature constituting a cycle as displayed in a deep-sea sediment core.

[4] The term was first brought to the attention of American readers by Antevs (1955, p. 324), who, however, did not utilize it.

In a set of provisional terms such as the one here offered it is not permissible to use any designation of rank belonging to standard geologic nomenclature. It is amply sufficient to refer informally to Neothermal time and the Anathermal, Hypsithermal, and Hypothermal phases thereof.

The period of rising temperature, attested by recession of glaciers and a slight rise of sea level, that has characterized the last two centuries is too brief and too close to us to justify a formal place in this terminology.

Those who feel that a provisional terminology cannot be substituted for standard geologic nomenclature, even those who consider the Recent to be a valid time unit (Morrison *et al.*, 1957), may still find useful the set of terms here proposed. It may be taken as a supplementary sequence emphasizing climatic progression, based upon stratigraphy as all sequences must be, differing in its subdivisions from a scheme in which climate is not directly concerned.

As a postscript to this discussion I submit a plea for uniformity in point of reference used in relation to number of C^{14} years before the present. The initial statement of age determinations is always in actual number of years elapsed. Some authors utilizing C^{14} data hold to this (B.P.), others convert to B.C.–A.D., and in certain papers both systems are used. The date of Christ's birth is of no geologic importance, and conversion to the system based upon it necessitates mental reconversion for most workers; archaeologists are perhaps the only important group for whom conversion to B.C.–A.D. is useful. It may be urged that the birth of Christ provides a fixed reference date, while the "present" is constantly moving onward. Considering the magnitude of probable error in C^{14} determinations, the approximate present will be a satisfactory reference datum for many years to come, and in future centuries the publication dates of our works will supply the means for correction. Universal use of B.P. in relation to C^{14} determinations is strongly urged.

References Cited

Ahlmann, H. W:son, 1953, Glacier variations and climatic fluctuations: Am. Geog. Soc., Memorial Lecture Series, III, 51 p.

American Commission on Stratigraphic Nomenclature, H. D. Hedberg, *Chairman*, 1952, Report 2-Nature, usage, and nomenclature of time-stratigraphic and geologic-time units: Am. Assn. Petr. Geol. Bull., v. 36, p. 1627–1638

Antevs, Ernst, 1948, The Great Basin, with emphasis on glacial and postglacial times: Univ. Utah Bull., v. 38, p. 168–191

——1953, Geochronology of the Deglacial and Neothermal ages: Jour. Geology, v. 61, p. 195–230

——1955, Geologic-climatic dating in the West: Am. Antiquity, v. 20, p. 317–335

Broecker, W. S., 1957, Evidence for a major climatic change close to 11,000 years B.P.: Geol. Soc. America Bull., v. 68, p. 1703–1704

Charlesworth, J. K., 1957, The Quaternary era, with special reference to its glaciation: London, Edwin Arnold, Ltd., 2 vols., 1700 p.

Chiarugi, A., 1936, Nuovo giorn. bot. ital., v. 43, p. 3

Coleman, A. P., 1929, Ice ages recent and ancient: New York, The Macmillan Co., 296 p.

Daly, R. A., 1934, The changing world of the Ice Age: New Haven, Yale Univ. Press, 271 p.

Deevey, E. S., and Flint, R. F., 1957, Postglacial Hypsithermal interval: Science, v. 125, p. 182–184

Emiliani, Cesare, 1955, Pleistocene temperatures: Jour. Geology, v. 63, p. 538–578

Flint, R. F., 1957, Glacial and Pleistocene geology: New York, John Wiley and Sons, Inc., 553 p.

Flint, R. F., and Deevey, E. S., 1951, Radiocarbon dating of late-Pleistocene events: Am. Jour. Sci., v. 249, p. 257–300

Harrison, A. E., 1956, Multiple glaciation since the Ice Age: Science, v. 124, p. 181–182

Judson, Sheldon, 1953, Geology of the San Jon site, eastern New Mexico: Smithson. Inst. Misc. Coll. 121, no. 1, p. 1–70

Just, Theodor, 1957, Postglacial vegetation of the North-Central United States (Abstract): Geol. Soc. America Bull., v. 68, p. 1895

Karlstrom, T. N. V., 1955, Late Pleistocene and Recent glacial chronology of south-central Alaska (Abstract): Geol. Soc. America Bull., v. 66, p. 1581–1582

——1956, The problem of the Cochrane in late Pleistocene chronology: U. S. Geol. Survey Bull., 1021-J, p. 303–331

Kinzl, H., 1932, Die grössten nacheiszeitlichen Gletschervorstoffe in den Schweizer Alpen und in der Mont Blanc-Gruppe: Zeitschr. Gletscherkunde, v. 20, p. 269–397

Matthes, F. E., 1939, Report of Committee on Glaciers, April, 1939: Am. Geophys. Union Trans. 1939, p. 518–523

——1940, Committee on Glaciers, 1939–1940: Am. Geophys. Union Trans. 1940, p. 396–406

——1942, Glaciers, p. 149–219 *in* Meinzer, O. E., *Editor*, Physics of the earth, IX, Hydrology, New York, McGraw-Hill Book Co., Inc., 712 p.

Morrison, R. B., Gilluly, James, Richmond, G. M., and Hunt, C. B., 1957, In behalf of the Recent: Am. Jour. Science, v. 255, p. 385–393

Von Post, L., 1931, Problems and working-lines in the post-Arctic history of Europe: 5th Intern. Bot. Cong., Cambridge, 1930, Rept. Proc., p. 48–54

——1946, The prospect for pollen analysis in the study of the earth's climatic history: New Phytologist, v. 45, p. 193–217

Smith, P. W., 1957, An analysis of post-Wisconsin biogeography of the Prairie Peninsula region based on distributional phenomena among terrestrial vertebrate populations: Ecology, v. 38, p. 205–218

Woldstedt, Paul, 1954, Das Eiszeitalter: Stuttgart, Ferd. Enke, 2d ed., 374 p.

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